THE LEWISIAN COMPLEX OF COLL, AND TIREE, ARGYLLSHIRE

by

Stephen A. Drury

Thesis submitted for the degree of Doctor of Philosophy under the general regulations of the University of Birmingham.

September 1970

UNIVERSITY OF
BIRMINGHAM

University of Birmingham Research Archive

e-theses repository

This unpublished thesis/dissertation is copyright of the author and/or third parties. The intellectual property rights of the author or third parties in respect of this work are as defined by The Copyright Designs and Patents Act 1988 or as modified by any successor legislation.

Any use made of information contained in this thesis/dissertation must be in accordance with that legislation and must be properly acknowledged. Further distribution or reproduction in any format is prohibited without the permission of the copyright holder.

BRITISH LIBRARY

IMAGING SERVICES NORTH

Boston Spa, Wetherby
West Yorkshire, LS23 7BQ
www.bl.uk

PAGE NUMBERING AS ORIGINAL

BRITISH LIBRARY

IMAGING SERVICES NORTH

Boston Spa, Wetherby
West Yorkshire, LS23 7BQ
www.bl.uk

BEST COPY AVAILABLE.

VARIABLE PRINT QUALITY

ACKNOWLEDGEMENTS

The work was carried out during the tenure of a University of Birmingham Research Scholarship, which is gratefully acknowledged. My supervisor was Dr. A.E. Wright who, with Dr. J. Tarney, read parts of the preliminary draft of this thesis.

Thanks are due to Dr. G.L. Hendry for his advice on x-ray analysis, Mr. L.W. Vaughan for his help with photography and Mr. A. Brown for his help with thin section making. A part of this thesis could not have been written but for the departmental library of computer programs largely written by Dr. A.C. Skinner.

I wish to thank the islanders of Coll, noteably Mr. and Mrs. J. MacFadyen of Glendyke, whose hospitality, warm hearted friendship and sincere interest provided the major encouragement to continue work in Birmingham with the prospect of escaping to Coll for each of three field seasons.

Finally, words cannot express the debt to my wife Joan, who, as well as typing this thesis, has had to bear my outpourings of woe, frustration and anger during its preparation.

SYNOPSIS

An area of Lewisian Gneiss some 150 sq km in extent is described, Coll in some detail, Tiree only partially. A history of deformation of at least seven phases is envisaged for the complex, although other indistinguishable phases occurred at an earlier stage in the history of the complex. Quartzo-felspathic gneisses, metasediments, basic, intermediate and ultramafic rocks, and early concordant granites show the varying effects of three metamorphic episodes. The earliest was at granulite facies and succeeded by amphibolite facies and greenschist facies metamorphism. Some tectonic phases may be correlated with metamorphic events. Certain geochemical evidence points to a pre-granulite facies lower grade metamorphism.

Many of the macroscopic features of the complex have been caused primarily by deformation. The types of structure in each phase were probably controlled by prevailing conditions, especially the presence or absence of hydrous fluids. Much evidence points to a chemical evolution of the complex perhaps analogous to a crustal layering of the metamorphic facies in basement rocks, and having a bearing on the processes active early in the history of this portion of the continental crust.

LIST OF CONTENTS

BINDING ERROR

Appendix 3: pages 194 - 203 follow page 214.

LIST OF PLATES

Plate 5 29

a. Isoclinally folded quartzose pegmatite, cutting unfolded migmatitic gneiss, deformed during D_3: 194598, Coll.

b. Sigmoidal boudins of concordant granite in granofels: Port na Luing, Coll.

c. Boudinaged metabasite, note flexure of foliation into cicatrices: Eilean Bhoramuill, Coll.

d. Boudinaged mafic band in migmatitic gneiss: 122520, Coll.

Plate 6 30

a. Rotated metabasite enclave in nebulitic gneiss: 177573, Coll.

b. Boudinaged hornblendic schlieren in nebulitic gneiss: 197534, Coll.

c. Phacoidal foliation and augen in deformed gneiss: Breachacha, Coll.

d. Type 1 interference structures, F_4 on F_2: 157563, Coll.

Plate 7 32

a. Type 2 interference structure, F_4 on F_2: 157563, Coll.

b. Type 3 interference structure, F_4 on F_2: 120519, Coll.

c. Type 2 interference structure, F_4 on F_2: 218620, Coll.

d. Type 3 interference structure, F_4 on F_2: Clabhach, Coll.

Plate 8 33

a. S-shaped asymmetrical fold: 162566, Coll.

b. Z-shaped F_4 fold with minor F_4 folds: 162568, Coll.

c. Disharmonic F_4 fold with axial plane structures: 220609, Coll.

d. Rotated F_4 folds and F_4 ptygmatic folds in remobilised concordant granitic veins: 220613, Coll.

b. Quartzo-felspathic augen in strongly deformed gneiss: Clabhach, Coll.

c. Enclaves of calc-silicate rock and basic material in nebulitic gneiss: Ballyhogh, Coll.

d. Boudinaged mafic schlieren in nebulitic gneiss: Friesland, Coll.

Plate 14 53

a. Diopside poikiloblast in diopside-gneiss, showing included quartz and felspar, and epitaxically intergrown and replacive hornblende.

b. Fractured, bent and granulated quartz and felspar in epidote-chlorite-gneiss. Patches of chlorite pseudomorphing biotite and hornblende are shown together with clusters of high relief epidote.

Plate 15 60

a. Plattung-type quartz with diagonal strain shadows, also sieved irregular garnet.

b. Deformed early garnet aligned with foliation defined by kyanite.

Plate 16 61

a. Trails of sillimanite needles within deformed garnet.

b. Sieved large garnets, unsieved small polygonal types, in garnet-biotite-gneiss.

Plate 17 65

a. Streaks of clinopyroxene and altered plagioclase with skeletal rims of garnet, separated by quartz with intergranular zoisite grains, in calc-silicate pod from gneiss.

b. Orthopyroxene mantled by clinopyroxene, separated by narrow quartz film from irregular garnet; bottom right. Garnet stringer cuts across brown hornblende; top left. Mafic granofels.

Plate 18 70

a. Biotite/quartz symplectite replacing garnet in mafic granofels.

b. Randomley oriented biotites in garnet-biotite-granofels.

Plate 19 75

a. Discordant mafic granofels cutting poorly foliated hypersthene-gneiss: Ben Hynish, Tiree.

b. Metabasite cutting massive hypersthene-gneiss: Ceann a' Mhara, Tiree.

c. Metabasite bands with discordant apophyses in well foliated hypersthene-gneisses with diffuse speckled garnet- and hornblende-rich granitic bodies: Hynish, Tiree.

d. Discordant metabasite cutting migmatitic gneiss: 243637, Coll.

Plate 20 76

a. Schlieren of metabasites in foliated migmatitic gneiss, some of which can be traced into dykes: Ceann a' Mhara, Tiree.

b. Banded garnet-amphibolite: Sorisdale, Coll.

c. Well developed agmatite: Pier road, Coll.

d. Rootless granitic bodies in metabasite, note the presence of hornblendite rims: 220610, Coll.

Plate 21 77

a. Agmatite in nebulitic gneiss of Rudha Hogh. The alignment and flattening of mafic fragments is due to D_3 deformation.

b. Unfoliated ultramafic rock with large orthopyroxene crystals: Bousd, Coll.

Plate 22 80

a. Garnet rimming orthopyroxene and ore, with intervening clinopyroxene and quartz adjacent to orthopyroxene.

b. Coarse, partially intergrown garnet and clinopyroxene from an "eye" in garnet-amphibolite. Note the replacement of clinopyroxene by green hornblende.

Plate 23 81

a. Margin of large sieved garnet with clear rim in garnet-amphibolite, showing "leached" zone around it consisting largely of quartz and felspar. Small orthopyroxene relics can be seen close to the leached zone.

b. Polygonal hornblende, plagioclase and quartz in recrystallised amphibolite.

Plate 24 84

a. Retrogressed garnets in amphibolite, pseudomorphed by fine grained aggregate of plagioclase, hornblende and ore.

b. Orthopyroxene with exsolved ore, amphibole and spinel.

Plate 25 85

a. Disequilibrium relations between orthopyroxene, pale hornblende and serpentinised olivine in hornblende-peridotite.

b. Polygonal, recrystallised matrix of hornblende in hornblende-peridotite.

Plate 26 86

a. Orthopyroxene with coexisting hornblende and plagioclase, in rim to hornblende-peridotite body at Bousd.

b. Felted actinolite-biotite rock in outermost rim to hornblende-peridotite body at Bousd.

Plate 27 90

a. Concordant granite veins in granofels at Sloc na Stuir.

b. Remobilised white granite with related concordant granitic veins in amphibolites and granofelses at Bagh na Coille.

Plate 28 92

a. Biotite network in biotite-pegmatite at Port na Luing.

b. Flinty crush rock with pumpellyite pseudomorphs after biotite.

LIST OF FIGURES

LIST OF TABLES

LIST OF MAPS (In rear pocket).

CHAPTER 1

INTRODUCTION

The islands of Coll and Tiree are numbered amongst the archipelago known as the Inner Hebrides, close to the West Coast of Scotland. They lie approximately 25 km west of Mull and 40 km south-east of the southermost of the Outer Hebrides, Barra (Fig. 1). Together, they form an incomplete outcrop of Lewisian rocks, about 45 km long perpendicular to the dominant WNW strike, having a total area of some 150 sq km.

In topography, the two islands are contrasted. Tiree is largely flat and featureless machair or grassed blown sand, with several outstanding hills. Coll has a rugged moorland scenery, similar to the Mainland Lewisian areas. Glacially rounded rock knobs up to 100 m high being separated inland by peat hags, and nearer to the shore by areas of machair and sand hills or bents. Shoreline exposures on both islands are excellent, as are the sandblasted outcrops in the dunes and machair. Inland exposures are sometimes lichen encrusted and this, coupled with the massive nature of much of the rock makes structural examination somewhat difficult, only the profiles of folds being generally discernable, the geometrical elements remaining obscure. Exposure estimated from aerial photographs of the main area of study, Coll, is about 25 per cent.

Selection of this particular area for study was encouraged initially by the Geological Survey maps (1925) which showed apparently interesting rock types and structures. Secondly by the isolation from the seemingly endless controversies of Mainland Lewisian studies, and the opportunity to study an almost virgin terrain of Lewisian gneiss outside the limits of the Mainland belts. Finally preliminary air photograph examination showed the good exposure and promising structure of Coll.

Preliminary field mapping of Coll on overlays to aerial photographs at approximately 1:10,000 scale was begun in the summer of 1968. Detailed structural mapping on scales up to 1:2,500 enabled a tectonic history to be deciphered to which were related metamorphic events as determined in the field and by routine petrography. Several hypotheses regarding the provenance and origin of certain rock types were proposed. These were to be tested by major and trace element analyses of carefully collected samples of rock, initially from Coll (see appendix). To extend the field

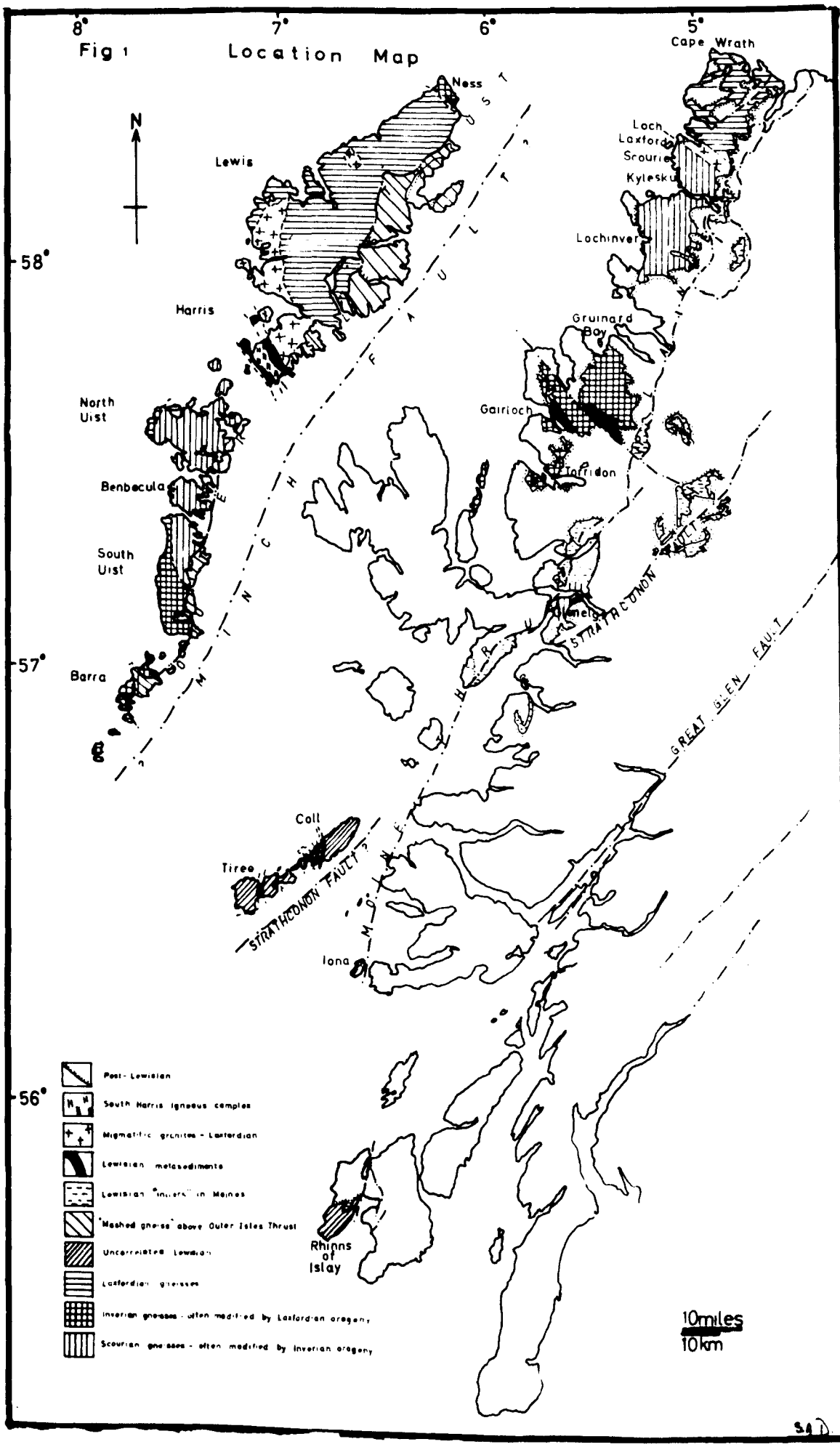


Fig 1 Location Map
8°
7°
6°
5°
58°
57°
56°
N
Ness
Lewis
Harris
North Uist
Benbecula
South Uist
Barra
Cape Wrath
Loch Laxford
Scourie
Kylesku
Lochinver
Gruinard Bay
Gairloch
Torridon
Coll
Tiree
Iona
Rhinns of Islay
MINCH
OUTER ISLES THRUST
MOINE THRUST
STRATHCONON FAULT
GREAT GLEN FAULT
Post-Lewisian
South Harris igneous complex
Migmatitic granites - Laxfordian
Lewisian metasediments
Lewisian "inliers" in Moines
"Meshed gneiss" above Outer Isles Thrust
Uncorrelated Lewisian
Laxfordian gneisses
Inverian gneisses - often modified by Laxfordian orogeny
Scourian gneisses - often modified by Inverian orogeny
10miles
10km

of study to some interesting occurrences on Tiree, a reconnaissance survey of the structure of this island was made together with the collection of samples for geochemical analysis.

In the course of the study some 600 petrographic slides were examined by the author, 400 of which were prepared by him, 200 being kindly lent by Dr. M.J. O'Hara. The latter were representative of metasediments, garnet-granofelses and metabasites collected by Dr. O'Hara on a short visit to Coll and Tiree in 1956. 210 samples were chemically analysed for 30 elements and oxides by the author using a Phillip's PW 1212 automatic X-ray fluorescence spectrometer and standard wet chemical methods. Minerals (felspars, biotites, garnets, amphiboles and pyroxenes) from analysed rocks were also examined using similar methods.

All grid references in the text refer to the National Grid (100 km square NM). All structural photographs were taken facing N, or down fold axe

Review of Previous Work.

Previous published work on the Lewisian of Scotland, except that dealing with the Lewisian inliers in the Moines, is reviewed briefly. More detailed discussion of some of the literature, where it has a bearing on the study undertaken, is presented later in the thesis.

An effort has been made to recount the research history of these Archaean metamorphic rocks and to show the present state of knowledge of the geological history of the Lewisian. Radiometric dates and geochemical data are summarised, showing the temporal subdivision and various genetic theories.

The location map (Fig. 1) is a compilation and interpretation of modern work on the Lewisian. It is by no means detailed, nor is it to be regarded as definitive.

Caledonian Foreland : Scottish Mainland.

In the latter half of the 19th century the Lewisian of the Mainland was mapped in some detail, the results being synthesised by Peach et al. (1907).

Officers of the Geological Survey divided the Lewisian into the "Fundamental Complex", and pre-Torridonian igneous intrusives cutting the Fund-

amental Complex.

The Fundamental Complex was said to consist mainly of gneisses, chemically and mineralogically similar to igneous rocks. Granitic and granodioritic gneisses contained bodies of basic and ultrabasic rocks. They were intercalated with rocks of sedimentary origin. At Gairloch, where metasediments were prominent Peach and Gunn thought them to be younger than the gneisses whilst Clough suggested they were the same age as the gneisses, which themselves were metamorphosed sediments, now unrecognisable as such.

The Survey noted a foliation and folds in the gneisses transgressed by basic and ultrabasic dykes, both gneisses and dykes having been deformed in places by pre-Torridonian movements. Where this post-dyke deformation was best developed, syn-orogenic migmatites had formed, and intrusive granites were found. Later brittle deformation was also recorded.

A threefold subdivision of the Mainland Lewisian into Northern, Central and Southern districts was proposed, based on the variation in magnitude of post-dyke deformation.

The Northern area, from Cape Wrath to Laxford Bridge was shown to consist mainly of hornblende- and biotite-gneisses in which the dykes were modified to concordant "epidiorite" bands as a result of strong deformation. A great development of granitic sills and dykes was noted.

In the Central district between Laxford and Loch Broom, the presence of large areas of pyroxene-gneisses, including bodies of basic and ultrabasic nature, was recognised. The dominantly flat lying foliation was folded gently by NE - SW axes. Those folds were earlier than vertical WNW trending basic dykes, themselves cut by E trending olivine-gabbros and picrites. Post-dyke movements, in the form of narrow zones of shearing and folding, modified the pyroxene-gneisses and dykes to hornblende-biotite-gneiss and hornblende-schist respectively.

The Southern region from Gruinard Bay to Torridon showed similarities to the Northern part, consisting largely of hornblende-biotite-gneiss. However it was noted that more recognisable dykes could be seen, and the post-dyke movements had not been so profound as in the Northern region.

The work of Sutton and Watson (1951 a) served to re-affirm the ideas of the Survey workers, but provided the first orogenic approach to the subdivision of the Lewisian. They showed that in two widely separated areas of Lewisian; between Scourie and Loch Laxford, early high grade pyroxene-gneisses with metasedimentary relics and ultrabasic bodies were

cut by a dolerite dyke swarm. The igneous intrusives were postulated to have been emplaced in one anorogenic episode associated with uplift and possibly further sedimentation, when the crust was rigid enough to fracture, and cool enough to give chilled margins to the dykes. A later period of orogenesis resulted in deformation and retrogressive metamorphism, and migmatisation of the pyroxene-gneisses, and dykes. The early high grade episode was termed Scourian, the dykes were called the Scourie dyke swarm, and the later orogenic episode was called Laxfordian. The scheme adopted for the two areas of study was applied to the rest of the Mainland. Like the Survey workers, Sutton and Watson used a threefold division of the Mainland Lewisian outcrops. The Central region was considered to be one of pyroxene-granulite facies rocks of the ancient Scourian complex, cut by the Scourie dyke swarm and only partly modified by the Laxfordian. The Northern region was seen as being made up of reworked Scourian gneisses on which a new NW trending foliation was superimposed during the Laxfordian. The gneisses' foliation and the basic dykes' margins were rotated into parallelism. Pyroxene-gneisses were entirely reconstituted and migmatised, biotite and hornblende becoming the ubiquitous ferromagnesian minerals. Large granite and pegmatite sheets were emplaced, microcline being strongly developed. Dykes became amphibolites difficult to separate from other basic gneisses (Sutton and Watson 1951 b). A similar pattern was recognised in the Southern region.

Later work (Sutton and Watson 1962) served to re-affirm their earlier conclusions.

Much recent work on the Lewisian has concentrated on study of the dykes and their relationships with structures and metamorphism, and of the ultrabasic bodies. O'Hara (1961) in studying the association of garnetiferous basic gneiss with ultrabasic rocks in pyroxene-gneisses near Scourie concluded that the former originated by reaction of ultrabasic intrusive dunites with country rock under granulite facies conditions. Bowes et al. (1964) regarded the banded ultrabasic masses such as that at Scourie, as being formed by tectonic fragmentation of large, initially stratiform basic-ultrabasic intrusions, with subsequent granulite facies metamorphism. Their conclusions were based upon the banded nature of the ultrabasic bodies, and the presence of structures similar to sedimentation features. O'Hara (1965) considered the evidence for sedimentation structures such as wedge and catenary bedding to be ambiguous and pointed out that the structures

were more probably produced by transposition of fold limbs. Bowes et al. (1966) presented chemical analyses showing chemical trends in the Lewisian ultrabasics similar to both stratiform and alpine-type intrusions and refuted the idea that the banding was a metasomatic manifestation.

Studies in central Assynt by Evans and Tarney (Evans 1964, Evans and Tarney 1964, Tarney 1963 and 1965) have shown that a period of folding producing a series of monoclines, such as the Lochinver antiform, preceded the intrusion of basic dykes, and to some extent guided them. Associated with the strong deformation was an almandine-amphibolite facies retrogression of the pyroxene-bearing gneisses. This orogenic episode was called the Inverian by Evans (1964), who defined the type area as the lower 3 miles of the River Inver where Lewisian gneisses with amphibolite facies assemblages are folded about NW trending axes. Regardless of the disputed nature of the type area, which contains a major Laxfordian structure, the Glen Canisp Shear Belt (see Bennison and Wright 1969, Fig. 3.3), the Inverian episode between the Scourian and Laxfordian has been widely accepted.

Detailed work on the relative ages, chemistry and mineralogy of Lewisian basic and ultrabasic dykes in Assynt by Tarney (1963, 1964) suggested that a series of dykes were successively emplaced in hot country rock during the later stages of the Inverian. O'Hara (1961, 1962) also discussed dyke intrusion in the Scourie region and concluded that dykes were emplaced under conditions of high temperature and pressure.

The existence of Laxfordian basic dykes has been postulated by a series of workers. Bowes and Khoury (1965) considered the younger of two dykes cutting Scourian gneisses north of Scourie to be Laxfordian. The early dyke was amphibolised whilst igneous textures remained in the later one. A series of basic dykes south of Gairloch was considered to be Laxfordian by Bowes and Ghaly (1964) and by Park (1964).

Much attention has been paid to that part of the Southern region around Gairloch and Loch Maree, where a series of Lewisian metasediments is present within the gneisses. Park (1964), working in the Gairloch area, recognised a longer structural sequence in an area of pyroxene-gneisses, the Ialltaig block, than in the surrounding amphibolite facies metasediments and associated basic rocks. He interpreted the Ialltaig block as being a fragment of Scourian basement tectonically emplaced in the cover of post-Scourian metasediments. He traced a gradation from the metasediments into quartzofelspathic gneisses, considering the latter to have been derived from the

sediments. The orogenic history of the metasediments and associated sheets was assigned to the Laxfordian. Since the South Sithean Mor dykes cut the gneisses of supposed sedimentary origin they were said to be Laxfordian, being earlier than late phase folding under amphibolite facies conditions. Sutton, in discussion to Park (1964), considered the gneisses at South Sithean Mor to be pre-Laxfordian in age, whilst the sediments were a younger series deposited on top of them and the South Sithean Mor dykes.

Applying his structural history of the Gairloch area further north Park (1965) found that between Gairloch and Poolewe gently dipping gneisses with relics of a granulite facies mineralogy could be ascribed to an Ialltaig, i.e. Scourian, age, and their foliation was cut by deformed basic dykes. Two sets of folds, correlated with separate phases at Gairloch, affect these gneisses. One of them, the Tollie antiform, had a steeply dipping SW limb in which the older foliation was seen to be transposed, obliterated and replaced by a newer, Laxfordian foliation.

In recent years the Glasgow School under the supervision of Dr. D.R. Bowes has produced a large amount of data for areas in the three Mainland districts collated by Bowes in 3 recent papers (1968 a, 1968 b, 1969). In these works overall correlations for the Lewisian of the Caledonian foreland and correlations with other areas of Pre-Cambrian rocks, in particular N. America and Scandanavia, have been propounded. Some of the main conclusions are listed here :

1) The orogenic approach to Lewisian history can be refined, elevating the three major Lewisian events - Scourian, Inverian and Laxfordian, to the status of separate orogenies, each probably being preceded by volcanic and sedimentary periods of deposition and including polyphase deformation and polymetamorphism. The present disposition of the Lewisian gneiss has been interpreted as the result of these superimposed orogenic belts.

2) The Central block of Scourian rocks was formed from a volcanogenic pile of dacites, andesites and basalts with subordinate sediments, termed the Kylesku Group (Khoury 1968 b).

3) The banded ultrabasic rocks lying within the Scourian foliation display original igneous layering and are tectonically emplaced fragments of an ultrabasic stratiform mass (Bowes et al. 1964).

4) The Scourian shows at least four separate tectonic phases (Khoury

1968 a, b).

5) The Inverian shows three folding phases before dyke intrusion. One set of dykes was deformed by an Inverian phase of folding, later sets were not.

6) Laxfordian shears deformed some dykes in the central region whilst some dykes cut shears (Bowes and Khoury 1965). Therefore both Inverian and Laxfordian dykes exist. Further evidence was cited from the Gairloch - Loch Tollie area (Bowes and Ghaly 1964, Park 1964, Ghaly 1966) and the area north of Laxford (Bowes 1969).

7) The gneisses north of Loch Laxford consists of migmatised sediments deposited after the Scourian and are not reworked Scourian gneisses. These metasediments were termed the Rhiconich Group (Bowes 1969). Sediments of the Loch Maree Series, including the Gairloch sediments, show gradational relationships with quartzo-felspathic gneisses, hence gneisses of the southern region were derived from sediments of post-Scourian but pre-Inverian, or possibly, post-Inverian, pre-Laxfordian age.

Outer Hebrides.

The basis for understanding the Lewisian of the Outer Hebrides has been the reconnaissance study by Jehu and Craig (1923-1934).

Dearnley (1962) described an overall scheme for the structure and metamorphism of the Outer Isles based on the hypotheses of Sutton and Watson (1951), and made a more detailed study of the Lewisian rocks of South Harris (Dearnley 1963). Like Sutton and Watson, Dearnley divided the Lewisian of the Outer Isles into three zones; a central block of Older Rocks from the Sound of Harris to central South Uist, a northern district comprised of Harris and Lewis and a southern district from central South Uist to Barra Head, both comprised of the Older Rocks reworked by a later orogeny. The whole was affected by very late movements resulting in the production of the Outer Isles Thrust.

Dearnley (1962) interpreted garnet-two pyroxene-basic bands cutting amphibolite facies gneisses in North Uist as representing the Scourie dyke swarm of the Mainland, and showing the existence of a pre-dyke Scourian complex. These dykes were equated with the metagabbros, anorthosites and tonalites of South Harris (Dearnley 1963) which were said to be the deformed remnants of a differentiated pluton. Since all these rock types showed

mineralogical affinities to the granulite facies, Dearnley suggested an early Laxfordian granulite facies metamorphism in the Outer Hebrides. He (1963) established zones of retrograde metamorphism in the rocks of South Harris, attributing them to a late Laxfordian amphibolite facies metamorphism. The granites and migmatites of North Harris were said to have formed during this stage. The remainder of the northern zone, being made up of hornblende-biotite-gneisses with some discordant but deformed and retrogressed basic intrusions was interpreted as being of Scourian origin, doubly reworked in the Laxfordian orogeny (Dearnley 1962). Similarly, Dearnley regarded the southern zone as being made up of reworked Scourian rocks. The arrangement of northern and southern Laxfordian zones about a central Scourian block suggested a similar, but spatially displaced, pattern in the Lewisian of the Outer Hebrides to that of the Mainland. Correlation of major Laxfordian structures across the Minch was postulated; for instance the NW-SE trending Seaforth antiform in Lewis was said to be a possible westerly extension of the Rhiconich antiform in Sutherland. The relative displacements of the belts and major Laxfordian structures was accounted for by a postulated post-Lewisian, sinistral wrench fault in the Minch now marked by a bathymetric trench.

Dearnley and Dunning (1968) ascribed several hornblende-gneisses with concordant pegmatites at localities on South Uist and Benbecula to a pre-Scourian period of formation. Agmatisation, and formation of the pegmatites was said to have occurred in the Scourian, whilst discordant basic dykes with granulite-facies cores and amphibolised margins were evidence of the double Laxfordian metamorphism.

Several workers in the Outer Hebrides have noted metasediments from the Lewisian, the best developed being on South Harris, described by Davidson (1943) and Dearnley (1963). Recently research workers from Imperial College, London discussed the widespread occurrence of metasediments throughout the Outer Hebrides (Watson 1968, Coward et al. 1969). Marbles, calc-gneisses, pure and impure quartzites, garnet-gneisses, anthophyllite-gneisses, aluminous gneisses, and various schists were found to occur as deformed belts, bands and pods on all scales within the predominant gneisses. Suggestions made by these authors are discussed later in the thesis.

Inner Hebrides.

Lewisian Gneiss is to be found on the islands of Islay, Iona, Coll and Tiree (see Fig. 1). Since the preliminary examinations by the Geological Survey of Scotland (1907, 1925, 1930) they have been ignored almost completely until recently. Several references by modern workers in the Lewisian to these localities have been founded apparently on the Survey's incomplete findings based on old-fashioned concepts of metamorphic rocks.

The Lewisian of Islay outcrops on the Rhinns peninsula over about 30 sq km. The Survey workers noted that porphyroblastic, acid hornblende-biotite-gneiss was cut by basic dykes. The latter sometimes maintained discordant relations and high grade assemblages; ortho- and clinopyroxene, brown hornblende and plagioclase. In other cases they had been deformed and retrogressed to green hornblende-plagioclase rocks. The whole suite was said to be characterised by large scale mylonitisation and crushing after the deformation of the dykes. The dominant mineral assemblages associated with this late shearing was epidote-sericite-quartz. It was found that some shearing was to be seen in the overlying Torridonian rocks and that the boundary was a shear zone. Thus the shearing was said to be post-Torridonian, at least in part.

About 3 sq km of Lewisian is to be seen on Iona where it is unconformably overlain by unmetamorphosed, but folded Torridonian. The Survey (1920) reported dominant acid hornblende-biotite-gneiss with concordant basic hornblende schlieren and concordant garnet-diopside-hypersthene-hornblende-plagioclase rock termed "eclogite". The strike was seen to be roughly NE and dips steep. Again the rocks were said to be characterised by late crushing and epidote veining. Pegmatites of a granitic nature were seen to be common. Many thin bands of undoubted sedimentary origin were found as thin concordant schlieren in the gneiss. Rock types reported were serpentine-marble, talc-marble, tremolite-marble, graphite schists, siliceous actinolite-schists, and garnet-magnetite rock. A metasedimentary origin was proposed for unfoliated "garnet-biotite-granulite", clinopyroxene-orthopyroxene-hornblende-plagioclase rock and orthopyroxene-hornblende rock, which might well be disputed now. A most unusual pegmatitic development of pure albite in an elongate mass was not ascribed to any particular mode of origin.

The only available works on the Lewisian of Coll and Tiree are the

Survey Memoir (1930), unpublished Geological Survey field slips of Tiree, Sinclair's unpublished thesis (1964) and Sinclair (1966).

The Survey (1930) indicated the predominance of "grey orthogneiss" on both Coll and Tiree with common, well defined belts of paragneiss and "dark orthogneiss". Zones of granitic gneiss of a well developed nature were described from both islands. Paragneisses examined, included various marbles, calc-silicate rocks, quartzites, magnetite bands and schists of garnetiferous, micaceous and graphitic type. A metasedimentary origin was ascribed to massive unfoliated "garnet-biotite-granulite" with subordinate orthopyroxene and amphibole, this rock accounting for about one third of the volume of paragneiss.

The "dark orthogneisses" were said to exhibit discordance with the grey gneiss only very rarely. Amphibolites, garnet-diopside-amphibolites and "pyroxene-granulites" were described, mostly from relatively narrow discontinuous bands. Large masses of dark hypersthene-diorite, or norite were said to occur on Ben Hough and Balephetrish Hill on Tiree, the latter containing marble xenoliths. The acid gneisses were grouped into three main types; grey hornblende-biotite-gneiss of granodioritic composition, pale hornblende-biotite-gneiss with indistinct foliation said to be deformed granite, and massive pyroxene-gneiss of uncertain affinities from Ben Hynish on Tiree. Numerous discordant pink biotite-pegmatites were recorded. All the above rock types were locally shattered and veined with green epidotic crush and black flinty crush. Comments on the structure of the gneisses were limited to the description of several major N-S trending isoclinal folds on Coll and Tiree affected by later cross folds trending ENE-WSW, and a large thrust-like crush zone carrying hypersthene-diorite over gneisses and metasediments on the west of Tiree.

Sinclair's (1964) account of the Lewisian of Tiree was largely one of the petrography of the gneisses. He recorded the following types of gneiss; dominant banded hornblende-biotite-gneisses of migmatitic nature, leucogranitic migmatite said to be recrystallised, contorted garnet-biotite-migmatite at Creagan Mora and Hynish, massive charnockitic acid gneiss on Ben Hynish, intermediate charnockite on Ben Hough. Basic rocks described were, pyroxene-granulites of Balephetrish and Balephuill, numerous thin, concordant amphibolites in the gneisses often garnet-bearing, and two cross-cutting bands on the west coast. Ultrabasic rocks were seen, noteably on Ceann a' Mhara, associated with marbles. The metasediment belts of Gott

Bay, Balephetrish and Loch a' Phuill were described petrographically in some detail. Sinclair accounted for the occurrence of juxtaposed granulite and amphibolite facies assemblages as being due to the varying availability of H_2O in a single, granulite facies metamorphism, a lack of amphibole/pyroxene relations in basic rocks being considered enough evidence for this. Basic bands were classed as resistant to H_2O diffusion whilst acid gneisses and metasediments were 'wet'. Pyroxenes were said to be developed at the expense of hornblendes. His mapping showed the form of the Ben Hough intermediate charnockitic to be lenticular as was the charnockitic migmatite of Ben Hynish, and he suggested that these might represent deformed intrusive bodies. No attempt was made to critically analyse the relationship between history of deformation and metamorphism.

Geochemistry of the Lewisian.

Several geochemical works on the Lewisian are outstanding, but in general little detailed chemical data has been published although numerous researchers have made large numbers of major and trace element analyses of Lewisian rocks from many areas.

Hitchon's work on the geochemistry of Lewisian pegmatites (1961) showed that the sequences of element fractionation were normal, magmas having been produced by anatexis of country rock at depth.

Park (1966) and Power and Park (1969) showed that concordant metabasite sheets in the Lewisian at Gairloch had a similarity in major and trace element compositions suggesting that they were all of a common origin, and magmatic. Strong chemical similarities to local basic dykes were interpreted as indicating co-genetic relationships. Bowes and Park (1966) studied the role of metamorphic differentiation in the formation of banded amphibolites from these basic sheets. They attempted to show that differential migration of Si, Na, K, Ca, Fe and Mg in hornblende-schists had produced striped rocks and ultimately hornblendic selvedges and felsic schlieren, under the influence of strong Laxfordian deformation.

Burns (1966) investigating chemical changes in Scourie dykes progressively deformed in the Laxfordian between Scourie and Laxford, found that little change in chemistry had occurred.

Bowes (1967) investigated the geochemistry of granitic rocks in three distinct structural settings in the Lewisian. He considered concordant,

potash-poor granites to have been extracted from the country rock by selective aqueous solution of quartz and albite and redeposited. Irregular diffuse permeations were probably formed by local partial melting and recrystallisation at low water fugacities. Large, massive potash-rich granites, obviously intrusive, were seen as the products of wholesale anatexis at depth.

Evans (1965) and Holland and Lambert (1969) noted the abnormally high K/Rb ratios, extremely low Rb content and unusually low Rb/Sr ratios of the Scourian gneisses. It was suggested, in both works, that it was very unlikely that the rocks contained normal crustal abundances of Rb for any length of time after their formation. They were therefore considered either to be from the base of the early crust, or that they represented normal crustal rocks which had been purged of elements, such as Rb, during a profound metamorphism shortly after their formation, equated with the Scourian granulite facies event.

Geochronology of the Lewisian.

Gilletti et al. (1961) made the first detailed study of the Lewisian. Finding a minimum 2400 myr date by the Rb/Sr method from a pegmatite in the Scourie area, they concluded that the Scourian orogeny occurred prior to this. Biotite ages were appreciably lower than felspar ages indicating a selective loss of radiogenic Sr^{87} during the Laxfordian reheating. Rb/Sr and some K/Ar dates from Laxfordian pegmatites and gneisses from the area north of Laxford and the Harris granite complex showed that Laxfordian metamorphism occurred between 1500 and 1600 myr ago. Dates from the Southern area included a biotite age of 1510±95 myr from a muscovite-biotite-gneiss on Gruinard Hill and 1525±45 myr from muscovite in a metasedimentary schist from Letterewe. In the Torridon area, low biotite ages (1300±30 and 1160±25 myr) were attributed to late shearing. The observed spread of dates was thought to be due to late Laxfordian reheating. These dates have provided only ambiguous possibilities of interpretation of dyke relations in the Southern area, Bowes, in discussion to Dearnley (1962), considered the dykes to be Laxfordian in age, as they cut gneisses giving a Laxfordian date. Dearnley, in reply, considered the dates to indicate Laxfordian recrystallisation of earlier gneisses.

From further work in the Gairloch area, Evans and Park (1965) suggested

that dates of; 1530 myr in an Ialltaig gneiss, 1400 myr in a metasediment, 1240 myr in an amphibolite sheet, and 1120 myr in a basic dyke from South Sithean Mor, indicated that early and main phase deformations occurred at least 1400 myr B.P.. However, they could have been earlier and perhaps of Inverian age. The South Sithean Mor dykes, later than the main phase of deformation and older than 1120 myr, were said to be possibly younger than 1530 myr, but perhaps older if the main phase deformation was earlier than indicated by its apparent age. These authors considered that a reheating episode at, or later than 1120 myr B.P. could account for the younger dates in the area.

The basic and ultrabasic dykes of Assynt have been dated by Evans and Tarney (1964) using the K/Ar method. These dates indicated that the majority of the Scourie dykes were intruded about 2190 myr ago, whilst some types of dykes were emplaced as late as 1950 myr ago. They found no dykes giving Laxfordian ages.

Evans (1965) investigated the geochronology of the Lewisian around Lochinver and proposed the following sequence;

i) Scourian granulite facies metamorphism culminated at least 2600 myr ago.

ii) Potassic pegmatites were formed 2250 myr ago.

iii) Inverian amphibolite facies metamorphism culminated about 2200 myr B.P..

iv) Immediately after the Inverian orogeny, dykes were emplaced, intrusion continuing for perhaps 300 myr under waning metamorphic conditions.

v) Parts of the area showed the effects of Laxfordian reheating dated at 1580 myr.

vi) Radiogenic strontium was purged from some metamorphic biotites by an event of uncertain character 1400 myr ago.

Evans also found a wide range of hornblende and biotite ages from 2100 to 1560 myr B.P. in amphibolite facies gneisses. He attributed them to varying argon retentivity during the Laxfordian reheating. A maximum Rb/Sr age of 2250 myr was given for the Inverian metamorphism, obtained from a pre-Inverian potassic pegmatite. The minimum age for the Inverian was that of a fresh picrite dyke cutting folded Inverian hornblende-gneiss, i.e. 2190 myr B.P..

In an attempt to find a minimum age for the formation of the Lewisian complex as a whole, Moorbath et al. (1967) made radiometric age determinations

on Lewisian fragments from basal Torridonian conglomerates. Most gave Laxfordian ages from 1798 to 1354 myr B.P., but an isotopic event 1150 myr ago was recognised. Whole rock ages for the Torridonian of 925 to 815 myr ago provided a lower datum for events contributing to the formation of the Lewisian. Moorbath et al. (1967) suggested that part of the Torridonian might have been derived from a Laxfordian or later supracrustal series containing metasediments and acid igneous rocks, which is not exposed now.

The geochronological study of the Lewisian, dealing with lead isotope data from Lewisian rocks in many settings by Moorbath et al. (1969), sheds new light on several problems regarding the history of the gneisses. They concluded that an event depleting the gneisses of uranium occurred at least 2900 myr ago and resulted in the present low U/Pb ratios in the Lewisian. This was correlated with the Scourian granulite facies, the younger maximum dates by Rb/Sr and K/Ar methods being attributed to later retrogressive events. The value of μ ($U^{238}/Pb^{204} = 8.68 \pm 0.09$) for the primary growth curve was comparable with the source region for modern oceanic lead, so that the rocks from which the Scourian complex was formed could not have had more normal, higher, U/Pb ratios for any length of time prior to uranium depletion. The results suggested that much of the Lewisian basement complex, including the parental material of the Laxfordian gneisses, was already in existence before 2900 myr ago.

Lambert (1970) has presented Rb/Sr data suggesting the Laxfordian orogeny was over by 1750 myr ago.

Lithological Terminology.

This short section provides a key to the terminology applied to groups of rock types found in the Lewisian Gneiss of Coll and Tiree used in the subsequent text. The arbitrary scheme is discussed with reference to various definitions found in geological literature.

Foliated quartzo-felspathic rocks.

Homogeneous, orthopyroxene-bearing quartzo-felspathic rocks are found between Balephuill and Hynish on Tiree. They are pale and exhibit faint mineralogical banding. The presence of orthopyroxene in such felsic metamorphic rocks might suggest the term "charnockite" be applied to them, as

has been the case with similar types from the Mainland (Sutton and Watson 1951a). "Charnockite" was first defined by Holland (1900) as a "quartz-felspar-hypersthene-iron ore rock". Later authors have applied the name to a wide variety of hypersthenic metamorphic rocks, ranging in composition from acid to ultrabasic (Parras 1958). Subramanian (1959), in a revision of the type area of Madras, suggested the term "charnockite" be restricted to what has previously been called "acid charnockite", preferably of magmatic origin, the intermediate, basic and ultrabasic divisions being regarded as alien to charnockites sensu stricto. The type charnockite, Job Charnock's tombstone in Madras, has been shown to be garnet-bearing. To avoid confusion, and because of the undoubtedly complex history of these rocks from Tiree, the term hypersthene-gneiss is preferred. "Gneiss" is used in the sense of Dietrich and Mehnert (1960); "a roughly or poorly foliated metamorphic rock, the approximate composition of which is indicated by suitable petrographic, mineralogic or even petrogenetic modifying terms".

Orthopyroxene- free quartzo-felspathic gneisses occupy the largest area on the islands. They are migmatites or migmatitic gneisses as defined by Dietrich and Mehnert (1960), being "macroscopically composite rocks that once consisted of geochemically mobile and immobile (or less mobile) parts manifesting themselves as leucocratic and melanocratic bands, or igneous-appearing and metamorphic materials". This definition is based only on field appearance, and "migmatite" may be applied to metamorphic rocks in which the "igneous-appearing" portion is a result of one of several possible processes. The dominant migmatite is hornblende-biotite-gneiss.

Within the hornblende-biotite-gneisses are belts up to 500 m across and narrow discontinuous bands of poorly foliated gneisses containing hornblende and biotite. In them are enclaves and schlieren of mafic, ultramafic and metasedimentary rocks, aligned in a parallel fashion, but irregularly distributed. In Dietrich and Mehnert's scheme (1960) these rocks could be termed nebulites or nebulitic gneisses. Associated with the nebulitic gneisses, where they abut metasedimentary rocks, are diopside-gneisses.

Chlorite-epidote-gneisses are to be found in association with later shear zones.

Rocks of possible sedimentary origin.

Rocks having textures and mineral assemblages visible in hand specimen that make their origin as metamorphosed sedimentary rocks a strong possibility are generally termed metasediments. Such rocks are widespread on Coll and Tiree as continuous variegated belts up to a km in width, impersistent bands up to 30 m across, and pods, lenses and enclaves within the quartzofelspathic gneisses up to 20 m in diameter but usually only of the order of a metre in size.

Rock types such as calc-silicate-quartzites, calc-silicate-calcite-quartz-mica-schist, calc-silicate-marbles, mica-quartzites, quartz-mica-schists and graphite-schists, are found in all scales of occurrence and are obviously metamorphosed sediments as are certain rocks intimately associated with them and demonstrably derived from them such as diopside-scapolite pods. A prominent magnetite band near Loch a' Phuill on Tiree is almost certainly sedimentary in origin.

Infrequently associated with the obvious metasediments, but fairly widespread as thin bands and pods in the migmatitic gneisses are garnet-biotite-gneisses and anthophyllite-gneisses. The garnet-biotite-gneisses sometimes carry Al_2SiO_5 polymorphs, and because of this are similar to rocks from other areas of high grade metamorphism designated metasediments of an aluminous nature. The anthophyllite-gneisses, because of the contrast of their mineralogy compared with the enclosing gneisses, were designated metasediments in the field.

Homogeneous mesocratic rocks.

Outcrops of varying size and extent of buff weathering granular rock carrying only sparse amounts of quartz, are widespread in the gneisses of Coll and Tiree. Although they often contain biotite they rarely show any distinct foliation or banding, even in hand specimen. They are, for purposes of field examination, completely homogeneous. The term "granulite" in the sense of Holmes (1920); "a granular metamorphic rock", might be applied, but because of the confusion of its international usage and its application as a genetic term in metamorphic facies nomenclature, "granulite" is not preferred. Goldsmith (1959) defines the term "granofels" as a medium to coarse grained granoblastic regionally metamorphosed rock,

without, or with only indistinct, fabric. Rocks of similar texture have been described from West Greenland by Berthelsen (1960), and called granofelses. In this thesis rocks with felsic minerals dominant or coequal with mafic minerals are called granofelses, prefixed by mineralogical modifying terms.

Mafic rocks.

Dark coloured, sometimes banded, poorly foliated to granular rocks commonly occur as relatively thin, sharp margined boudinaged bands and pods in the gneisses and metasediments. Several prominent bands up to 500 m wide are present. Ferromagnesian minerals predominate in these rocks. Similar mafic metamorphic rocks at Gairloch have been called "basites" by Park (1964). Mafic rocks in Connemara have been called "amphibolites" by Leake (1964). In this thesis mafic metamorphic rocks are termed metabasites, as the name has no genetic connotations and the complete mineral assemblages cannot be discerned in the field. Where the mineral assemblage is known from microscopic examination, as in Chapter 3, the scheme of nomenclature devised for basic metamorphites by Berthelsen (1960) is adopted.

Ultramafic rocks.

Rocks devoid of felsic minerals in hand specimen are rare on Coll and Tiree. They may consist of aggregates of assorted mafic minerals or be monomineralic, in either case they have been called simply ultramafic rocks in the field.

Granitoid rocks.

This heading covers a multiplicity of granitic and granodioritic rocks of varied habit. Several types, distinct in setting and relative age have been distinguished. They are termed granites or pegmatites depending on their grain size, with qualifying prefixes depending on their distinguishing features.

Mylonitic rocks.

Veins and bands of aphanitic mylonitic rock are widespread. They are termed epidotic crush rock or flinty crush rock, depending on their appearance and mineral content.

CHAPTER 2

STRUCTURAL HISTORY

A lack of common, penetrative linear fabrics or of discordant relationships between intrusive rocks and the deformation phases seen, severely limits the enumeration and correlation of tectonic events. To some extent the lack of critical events has been overcome by rigorous examination of the effects of each phase on earlier structures, such as the refolding of folds and fold features, the folding of planar fabrics produced in different episodes, and the production of foliation in previously homogeneous materials.

Early structures, possibly representative of several phases of deformation, are seen in granulite facies gneisses in the western parts of Tiree. They are older than the age of intrusion of basic and intermediate igneous rocks. Since then the complex has been subjected to several phases of deformation, each manifesting itself in a different manner. Most are characterised by folds of distinct style, and in each phase contrasted lithologies were affected in different ways depending on their competence, itself a function of their composition and the prevailing metamorphic conditions. The rheological state of the complex as a whole is suggested to have varied during its tectonic evolution. The tectonic events are regarded as periods of time, not necessarily mutually exclusive, which may be used as a relative chronological scale to subdivide the period after the emplacement of igneous rocks up to the close of tectonism. Various metamorphic and igneous events may be correlated with this scheme. Each phase is referred to as D_n, n being the position in the sequence, as an aid to clarity in the text. The identified events and their associated features are summarised below:

Early structures.

D_1: Foliation S_1 and intrafolial folds F_1

D_2: Isoclinal folds F_2

D_3: Flattening of S_1 to form S_3 with lineation L_3

D_4: Asymmetrical folds F_4

D_5: Monoclinal folds F_5

D_6: Open folds and shear belts F_6, foliation S_6 in shear belts.

D_7: Brittle deformation.

Apart from the earliest structures, most of the descriptions and discussions contained in this chapter refer to observations on Coll, where the

widest range of relationships was seen, and detailed field mapping possible because of good exposure. Those structures seen on Tiree post-dating intrusion were in agreement with relationships on Coll and it seems reasonable to suggest that the gneisses of Tiree are not markedly different except that the effects of the deformation phases decrease westwards.

Early structures.

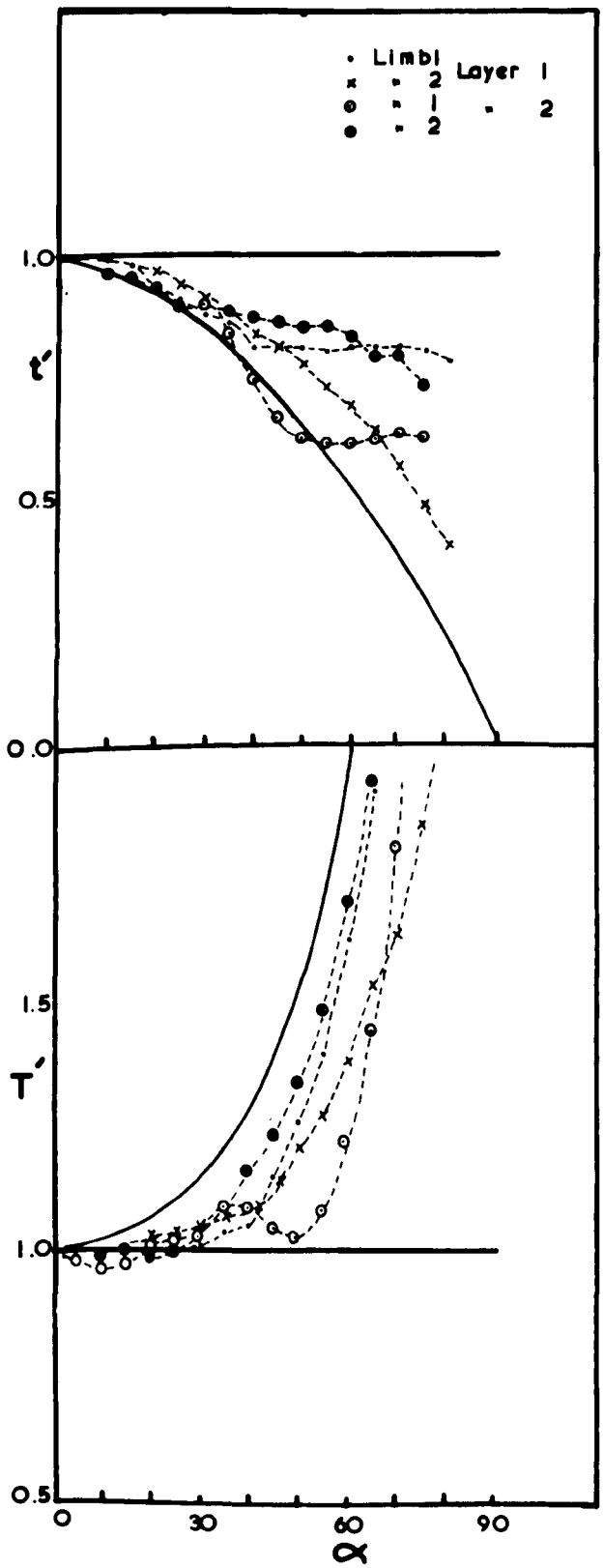
The earliest recognisable structures are seen in rocks showing affinities to the granulite facies, examined between Hynish Bay and Traigh nan Gilean on Tiree (see Map 3). Hypersthene-gneisses have a foliation defined by mineral banding in which lies a linear fabric defined by elongate aggregates of ferromagnesian minerals. The foliation contains isoclinal folds with limbs and axial surfaces parallel to the banding, and occurring as isolated, and probably dislocated, single closures. They are sparse and indistinct, probably having formed by extreme deformation of earlier folds, at the same time as the banding. The foliation is cut by unfoliated metabasite and granofels sheets (Plates 19a and c) which also show granulite facies assemblages (see Chapter 3 p.67). This shows that the earliest structures in the hypersthene-gneisses were formed prior to intrusion and granulite facies metamorphism. The effects of subsequent tectonic events on these structures was not observed, apart from metabasite dykes cutting massive hypersthene-gneisses at Traigh nan Gilean having been rotated into concordance in well foliated gneisses on the west shore of Cean a' Mhara (Plates 19b and 20a). The progressive rotation of the dykes is demonstrable.

The remaining tectonic events seen on Coll and Tiree occurred after the intrusions. In general the foliation is highly inclined on Tiree, usually vertical, and this steep belt continues on Coll to about the line of the Hogh Bay - Breachacha metasediment belt. The strike of the steep belt is variable. Eastwards from the Hogh Bay - Breachacha belt, dips are much shallower, being contained mainly in a quadrant from NE to SE and the overall structure of the archipelago is that of a large refolded monoclinal fold with a large angle between the limbs.

D_1: Formation of foliation S_1 and intrafolial folds F_1.

Extremely tight isoclinal fold closures, with limbs and axial planes parallel to the foliation, are quite common in migmatitic gneisses, especially in the steep southerly dipping gneisses of Caolas Ban on Coll. The inter-limb angle is invariably less than 30° and often very small. Using the classification of Fleuty (1964), the folds can be described as tight or isoclinal. They are defined by the thinner compositional bands and fold a foliation, but are contained by the thicker bands and parallel the dominant foliation. They are therefore termed intrafolial folds. Their size varies from a few centimetres to several metres, being typically from 10 to 50 cm.

The style of intrafolial folds is very varied, partly depending on the composition of the rocks containing them. They may occur as single hinges or in pairs, and as rootless closures in which the limbs are apparently sheared out (Plate 1a). Complex folds are present and may be harmonic or disharmonic in style (Ramsay 1967, p. 419) (Plate 1b). The hinges' angularity varies independently of gneiss composition, from rounded (Plate 1d) to very sharp (Plate 1c), the most angular folds, being those with extremely attenuated limbs. Most F_1 intrafolial folds can be assigned to classes 1C, 2 or 3 of Ramsay (1967), many being of type 3, with strongly attenuated limbs and the rate of change of slope of the outer fold arc exceeding that of the inner. A peculiar intrafolial fold occurring in acid gneiss enclosed by a thick granofels body at Eilean Ghreasamuil on Tiree has been geometrically analysed (Plate 1d Figs. 2a and b). It shows a geometry of class 1C near the hinges, but dip isogons in the limbs tend to become perpendicular to the folded surfaces, and the plots of t' (the relative orthogonal thickness) and T' α (the relative thickness of a folded layer parallel to the axial surface) against α (the angle of inclination of the folded surfaces) show a trend parallel to those of class 1B, i.e. parallel folds (see Ramsay, 1967 p. 360-369). Plate 1a shows intrafolial folds in banded migmatitic gneisses with a thin concordant metabasite band. Whilst the gneiss is foliated and folded, the metabasite is massive. The foliation and folds are cut by a white quartzose pegmatite. In all localities, metabasites and granofelses are devoid of F_1 intrafolial folds. This fact may well have a bearing on the origin of the intrafolial folds. Three possibilities exist for their formation. Firstly, they may have existed as some form of folding prior



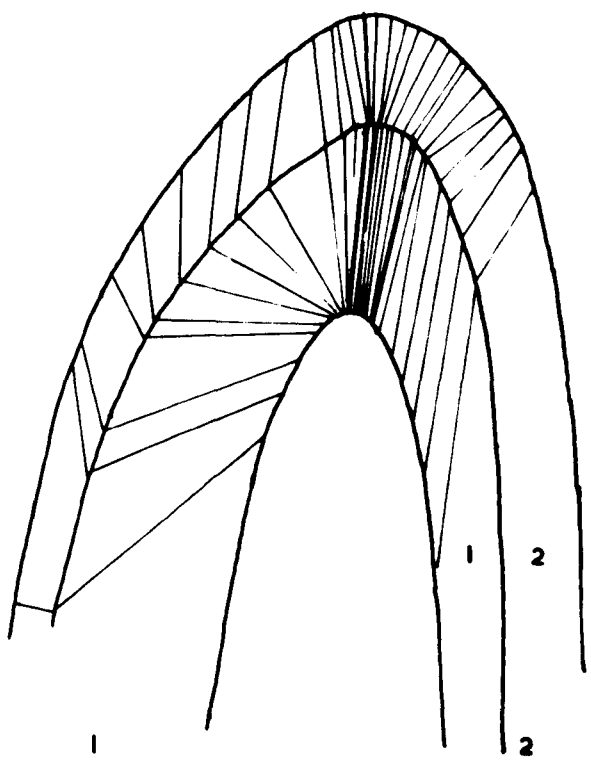


Fig.2: Graphs of t' andT' against α for F_1 intrafolial fold at gneiss/mafic granofels interface, and profile with constructed dip isogons: Eilean Ghreasamuil, Tiree.

PLATE 1

Plate 1

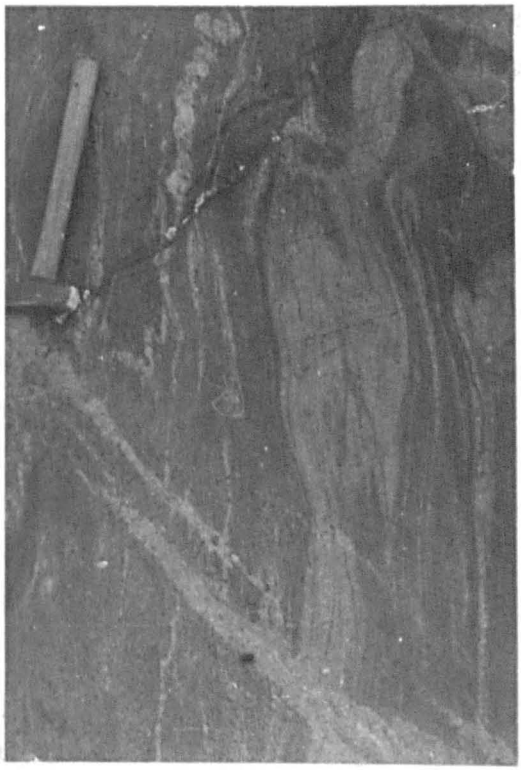
a. F_1 intrafolial fold in migmatitic gneiss; note unfoliated metabasite band and discordant quartzose-pegmatite: Traigh nan Gilean, Coll.

b. Disharmonic F_1 intrafolial folds in intermediate migmatitic gneiss: Caolas Ban, Coll.

c. F_1 intrafolial fold in mafic band: Caolas Ban, Coll.

d. F_1 intrafolial fold at gneiss/mafic granofels junction: Eilean Ghreasamuill, Tiree.

a



b

c

d

to metabasite and granofels intrusion, being subsequently deformed and transposed, the intrusions being rotated into parallelism. They may have formed after intrusion, deformation manifesting itself as folds and an enclosing foliation in the acid gneisses whilst the more competent intrusions were strained homogeneously, without folding. This would seem to be negated by the fact that intrafolial folds exist in mafic bands in the gneiss (Plate 1d), differing little in composition from the sharp margined metabasites. Also, Plate 1a shows an unfoliated metabasite adjacent to gneiss with F_1 folds. Finally, although F_2 isoclinal folds can be seen to fold F_1 intrafolial folds in a few cases (Plate 2a), some folds of intrafolial habit, especially in gneisses deformed by post-F_2 deformation phases, may represent deformed F_2 isoclines.

Intrafolial folds have been described from the Lewisian by Wheatley (1969) and Sheraton (1970), and Flinn has used the term "internal folds" to describe tight folds in the Shetlands formed by inhomogeneous flow of beds with little disturbance of limiting bedding planes. The form of the folds suggest that the containing rocks were in an incompetent state during deformation. Intrafolial folds may have formed by transposition of pre-existing folds, or as suggested by Turner and Weiss (1963 p. 92), transposition of a pre-existing banding or layering may produce tightly appressed folds with axial planes sub-parallel to the secondary foliation.

It proved impossible to evaluate axial trends of intrafolial folds in the field, and no linear structure was noted to be associated with them.

The foliation containing F_1 intrafolial folds is probably the product of the deformation producing the fold forms. It is described in detail in Chapter 3 in the section dealing with the migmatitic gneisses. To some extent the mineral banding in these gneisses existed before D_1 as shown by the presence of a foliation preserved around F_1 fold hinges. However it is reasonable to suggest that it is a new planar fabric, albeit produced by transposition and metamorphic reworking of an earlier gneissose banding. In the migmatitic gneisses, distinct metabasite bands are thoroughly concordant, only being distinguished from mafic segregation banding in the gneiss by their contrast in thickness to the mafic bands, sharp margins and the absence of intrafolial folds. The S_1 foliation is variably developed in metabasite bodies, which may range from fabric-free to strongly banded. Three plausible hypotheses as to their origin are recognised. Firstly that they were present in the complex at an early stage. They might rep-

resent intrusive rocks, either sills or rotated dykes of the same age as those of proven age on Tiree, or they might be later pre-F_2 intrusives. Similar origins may be postulated for the unfoliated granofels bands and bodies. The problem is returned to later in the text.

D_2: Isoclinal folding F_2.

Folds of a close to isoclinal nature are found on Coll and Tiree as major structures of a wavelength up to one kilometre, intermediate structures which may be mapped on a scale of about 1:10000, and small folds which are represented on Map 2 by symbols showing their orientations.

Major F_2 isoclinal fold axial traces are shown on Maps 2 and 3. Between Brock and Fhadamuill on Tiree, the axial traces of folds correlated with F_2 on Coll are folded about two successive major fold axes, the last being an open, neutral fold trending E-W.

An E-W trending F_2 isoclinal fold between Traigh Halum and Caolas Ban on Coll is folded by a N-S trending F_4 asymmetrical fold from Traigh Halum to the vicinity of Eilean Bhoramuill. The crest of the fold can be seen at 110530 where it is an antiform with a horizontal axis. Variation in attitude and axial plunge of this fold cannot be evaluated because of the lack of exposure in sand dunes behind Traigh Halum.

A marked major F_2 isoclinal fold with an axial trace trending N-S, is shown on Map 2 from west of Totronald to Port na Luing. At Port na Luing, its hinge zone is described by a granofels band and a metabasite body. In the migmatitic gneisses contained by the granofels band, the hinge zone is not seen, but the axial trace can be mapped using the symmetry of intermediate and small scale parasitic folds (see Map 2). Foliation measurements in the core of this fold do not indicate any closure, but in general M-shaped folds occur along the axial trace, S-shaped folds west of it and Z-shaped folds on the eastern limb. At Port na Luing, the attitude of this fold seems to be that of a steep southerly plunging, steeply inclined antiform, in the sense defined by Fleuty (1964), whose axial surface dips at a high angle to the east, parallel to the foliation. Because of the highly deformed nature of the gneisses in the core, the attitude and geometry of the major structure cannot be directly determined. However measurements on intermediate and small scale folds in the gneisses, as plotted on Map 2 and in Fig. 5 , suggest that the inclination of the axis

is highly variable, as is possible with steeply inclined folds (Fleuty 1964). Map 1 shows that the granofels horizon is thicker by possibly a factor of three in the hinge zone as compared with the sub-parallel limbs. This suggests that in this particular rock type, the fold approaches a similar style, being definitely in one of classes 1B, 2 or 3 of Ramsay (1967, p. 365-9). No complementary major F_2 closures are exposed with close relations to this one and it may be a rootless fold, akin to a megascopic intrafolial fold. However, the presence of a thin central band of highly deformed hornblende-biotite-gneiss in the Hogh Bay - Breachacha metasediment belt (Map 1) and the unusually large width of the Feall Bay - Crossapol Bay metasediments, 1.5 km by inference, may indicate that it is bounded by two very attenuated synforms containing metasediments.

No evidence of major F_2 folding was found in the largely nebulitic gneisses from Rudha Hogh to Fasagh to account for possible repetition of metasediments in the Clabhach - Gorton belt. This belt is now underlain and covered by gneisses, and two hypotheses could account for this attitude. Either distinctive sedimentary lithologies were originally interlayered with the parental material of the migmatitic gneisses, or this belt of metasediments was tectonically emplaced at some stage during the orogenic history of the complex. If the first case were true then the complex as a whole must have had a common time of origin, whilst the second could also be accounted for by a further possibility, that sediments were laid down on an earlier gneiss complex.

Isolated, tight F_2closures are outlined in several localities by metabasite horizons, noteably at 156563, near Totronald, and SE of Toraston. They are apparently rootless and only recognisable because of the outcrop form of a distinctive lithology.

Intermediate and small scale F_2 folds are considered together, and are generally the only representatives of this tectonic phase, occurring at wide intervals in well banded and foliated gneisses which are largely devoid of folding. However they show a marked concentration on Coll along the hinge zone of the major structure from Port na Luing northwards, where they are seen in several different rock types. On average, F_2 folds have a wavelength of a few metres and an amplitude several times larger. The size varies with rock type, being largest in well banded migmatitic gneisses, and smallest in concordant granitic veins in granofelses, although there is

a wide range in all lithologies.

In the migmatitic gneisses, F_2 folds generally have interlimb angles from about 60° to 0°, and may be classified as close, tight or isoclinal folds in the sense of Fleuty (1964). Many of the larger folds, such as that shown in Plate 2b from the axial zone of the Port na Luing - Totronald fold, show a large variation in interlimb angle along the axial surface. In this intermediate sized synform the layers at high structural levels in the fold have an interlimb angle of about 100°, and the fold becomes progressively tighter, so that the lowest exposed part has a tight to isoclinal form. There is no evidence in this case for the tightness being affected by different compositions of folded layers.

F_2 axial planes generally dip parallel to the regional foliation, and the folds vary in attitude from steeply inclined to upright in the Port na Luing - Totronald fold (Fig. 5), to recumbent in flat lying gneisses (Plate 2c).

In the outcrops with common folds of this generation there is a marked variation in fold style, complementary antiformal and synformal closures showing contrasted outlines in the hinge zone. Plate 2d shows a very tight synform where the curvature of the limbs increases to a maximum at one hinge. The adjacent antiform shows an unusual multiple-hinged outline in the outer part of the fold, the constructed axial surface in the core bifurcating, so that the outer part has a rectangular shape. The change from single- to multiple-hinged folds is seen in both antiforms and synforms, often in the same fold, and seems to be the result of compensation during some stage in the development of the present fold shape. This feature, coupled with the variation in interlimb angle in one fold, shows the partly disharmonic nature of some of the F_2 folds. The limbs of the folds where they have this box-like nature are highly attenuated and this might suggest they reached their present form in a fashion analogous to that of true box-folds, where limb deformation rather than limb faulting enabled the hinge region to be moved to allow for the accomodation of material in the fold core (Ramsay 1967, p. 421).

Angularity of the hinges is extremely variable, the most rounded being in the double-hinged parts of folds and the most angular being in the tightest folds in the migmatitic gneisses.

In some areas of migmatitic gneiss as at 208514 F_2 folds show an intrafolial habit, very tight closures, often partly double-hinged being separated

PLATE 2

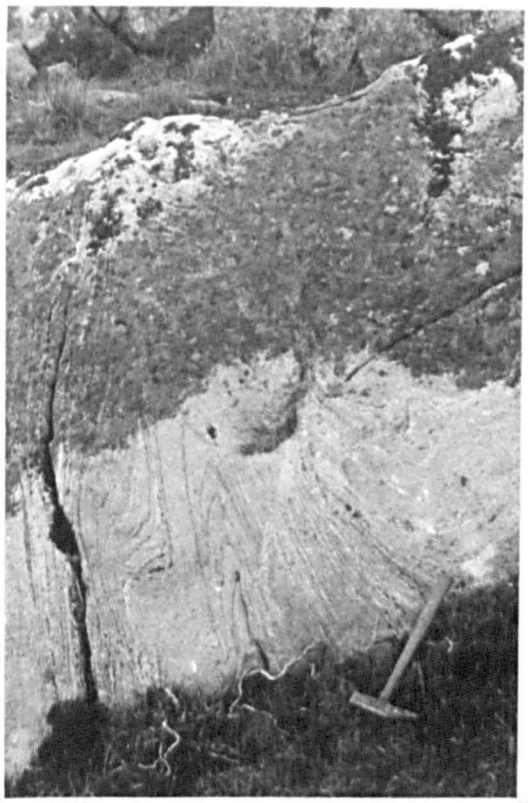
Plate 2

a. F_1 intrafolial fold refolded by F_2 isocline, itself probably deformed during D_3: 156560, Coll.

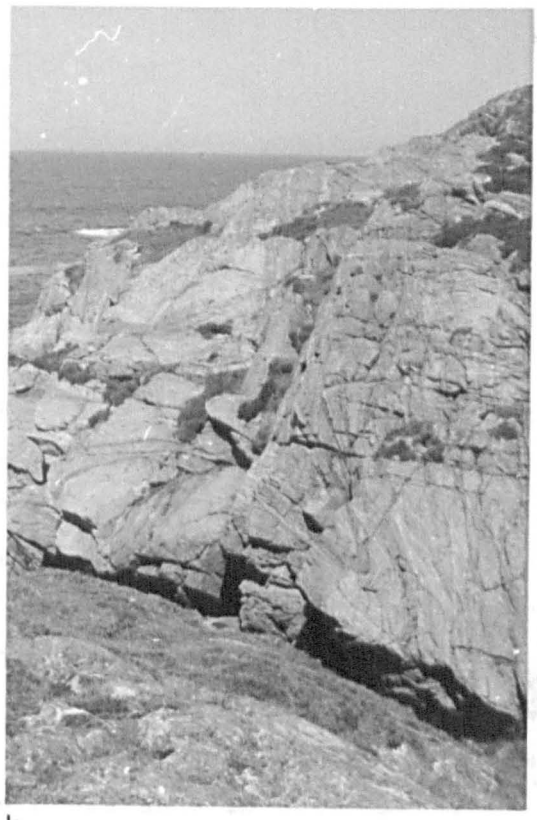
b. F_2 isoclinal fold showing change in interlimb angle along its axial surface: 157567, Coll.

c. Recumbent F_2 isoclinal fold: 223615, Coll.

d. Double-hinged F_2 antiform and tightly appressed synform: 156564, Coll.



a



b

c

d

from complementary closures by relatively thin parallel belts of highly attenuated gneiss, probably representing strongly deformed fold limbs. Although distinctive marker horizons in the folds are absent, it appears that isoclinal fold closures have been transposed relative to one another by movement parallel to their highly deformed limbs (Plate 3a).

Where F_2 isoclinal folds affect thin metabasite bands and mafic bands in the migmatitic gneisses, no apparent difference to the style in acid material is seen. However, where small F_2 folds are found in thick metabasites, as at Port na Luing, the folds have approximately equal wavelengths and amplitudes. They are much more regular in outline, and never with an interlimb angle less than 30°, being classed as close to tight folds. The hinges are consistently well rounded, and the limbs never show such a great degree of attenuation as is seen in the migmatitic gneisses (Plate 3b). At Port na Luing these folds are upright, and the limbs make a high angle with the axial planes. They are seen to be parasitic folds developed about the axial trace of the large F_2 fold whose hinge zone is seen there.

Isoclinal F_2 folds are found in concordant, often garnetiferous, granitic veins and bands in the granofelses. Large bodies of this white material have the shape of isolated isoclinal fold hinges, when carefully examined. Around the hinges of these folds, the boundary between granite and granofels frequently shows a highly irregular lobate form (Plate 3c). Rounded excrescences of granite protrude into the granofels, with a large wavelength and small amplitude. They are separated by infolds of granofels into granite with a very tight and angular form, short wavelength and relatively large amplitude. These fold forms are similar to those produced by progressive compressive strain of a single plane separating two homogeneous substances of differing viscosity, as illustrated by Ramsay (1967, Fig. 7-42, p. 383). The relative effects on the granofels-granite boundary suggests that during the formation of F_2 isoclinal folds, physical conditions were such that the viscosity of the granite exceeded that of the granofels. The thin granitic veins display complex elasticas-like folds with a negative interlimb angle in unfoliated granofels (Plate 3c). Such folds are often termed ptygmatic folds. If the initial wavelength/thickness ratio in a folded layer of higher viscosity than the surrounding medium is initially large, then the compressive strain which can be taken up by folding is greater than the maximum of 36 per cent shortening in a concentrically folded layer, and irregular ptygmatic folds are formed. Unlike many ptygmatic folds

PLATE 3

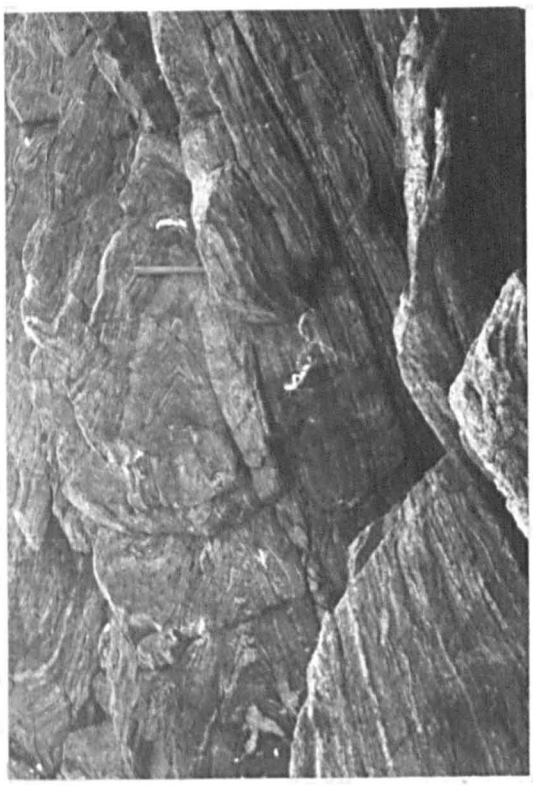
Plate 3

a. Recumbent F_2 fold with highly deformed lower limb: 211546, Coll.

b. F_2 fold in metabasite: Port na Luing, Coll.

c. F_2 viscosity contrast structures in granofels and concordant granitic vein: Port na Luing, Coll.

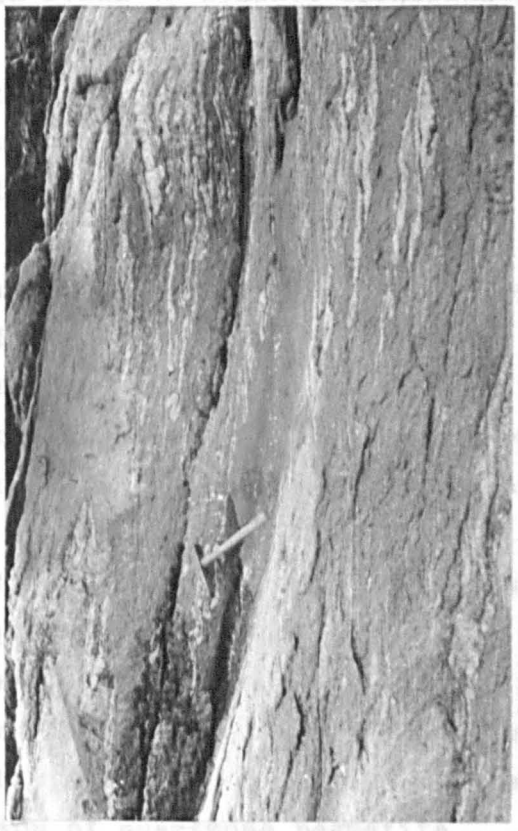
d. Deformed ptygmatic F_2 folds in granitic veins in granofels: 207543, Coll.



a

b

c



d

(Ramsay 1967, p. 111) those in the granofels do not have a concentric form and the thickness of the folded layer varies wildly around individual folds (Plate 3c). In some localities, noteably at 207543, ptygmatic folds in granitic veins within a granofels band are seen to have been transposed parallel to the regional foliation and now show dislocated habit and intrafolial style (Plate 3d).

In metasedimentary rocks, F_2 folds are quite common. Angular, upright, southerly plunging isoclinal folds are seen in the quartzites of Ben Feall. At Sloc na Stuir, a thin intercalation of calc-silicate rocks in a thick granofels band contains numerous irregular isoclinal folds, often of an irregular nature with multiple hinges and non-planar axial surfaces (Plate 4d).

Fold profiles with constructed dip isogons and graphs of t' and T' against α, are given in Figs. 3 and 4 for F_2 folds from different lithologies. Fold 20 (Figs. 3 and 4) has dip isogons and trends of t' and T' whose variation with α show that in layer 1 the fold is close to class 2 throughout. Layer 2 has features of classes 3 in limb 1 and 1C in limb 2, at high values of α. Layer 3 follows class 1C closely throughout.

The profile of fold 30 (Plate 4b and Figs. 3 and 4) from migmatitic gneiss, is taken from a multiple hinged fold, and varies along its axial surfaces. Both layers 1 and 2 have features of class 3 (where the slope change of the outer arcs exceed those of the inner) in the hinge zone, changing to class 1C in both limbs.

Fold 7 (Plate 3b, Figs. 3 and 4) from a metabasite body shows almost parallel dip isogons and the variations of t' and T' with α show that it closely approaches class 2 or a similar fold form.

Fold 11 (Plate 4a, Figs. 3 and 4) from a calc-silicate horizon shows regular variation in dip isogon inclination, t' and T' with α, indicative of its being class 1C.

Two possibilities might account for the present styles of F_2 folds; either they were produced by a single progressive deformation, or have a double history of evolution. In both cases, the modes of deformation could have been variable. Evidence presented in the section dealing with the D_3 phase strongly suggests that two phases contributed to their present geometry, separated by a period of intrusion of quartzose pegmatite bodies.

PLATE 4

Plate 4

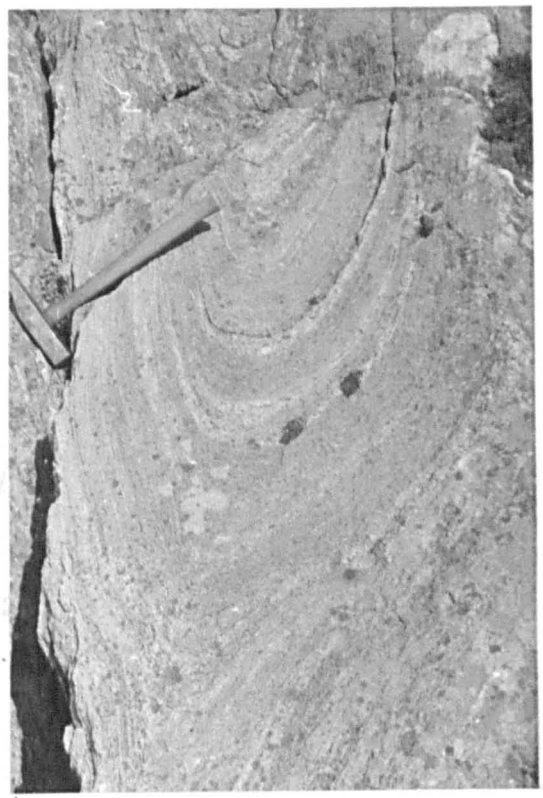
a. Multiple-hinged F_2 fold in calc-silicate rock: 157569, Coll.

b. F_2 fold of variable geometry in migmatitic gneiss: 157567, Coll.

c. Eye-like structures due to variation in F_2 axial plunges: 156555, Coll.

d. Discordant quartzose pegmatite, cutting F_2 axial surface, and itself isoclinally folded: 208543, Coll.

a

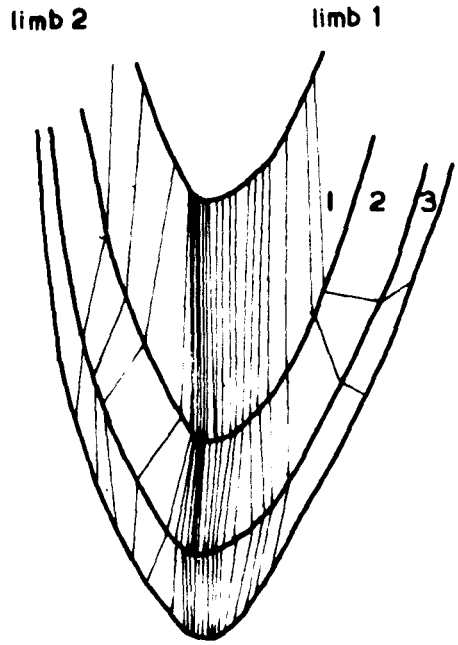


b

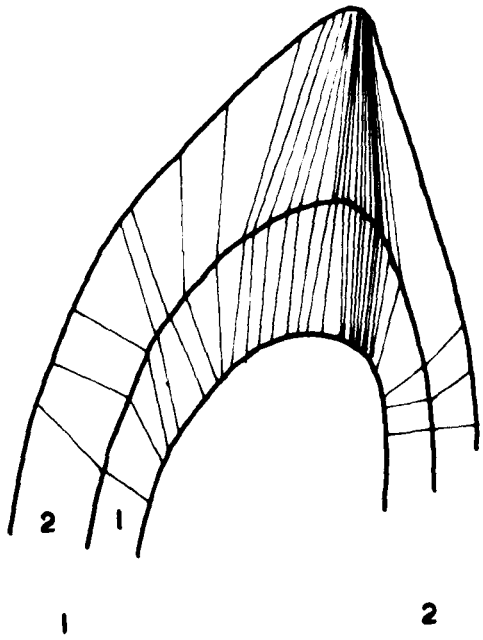
c

d

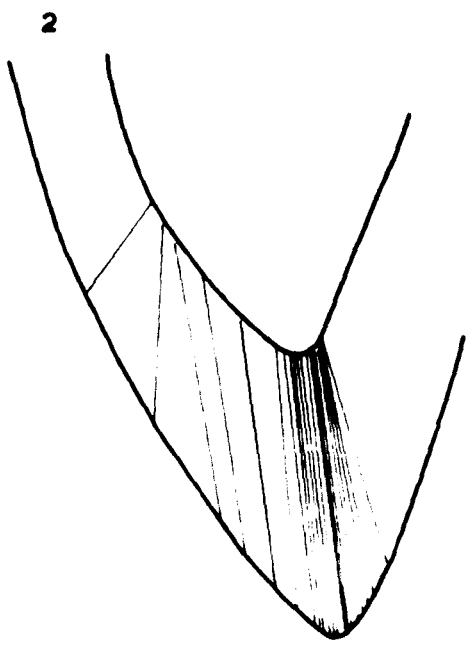
Fig 3 F_2 fold profiles with dip isogons



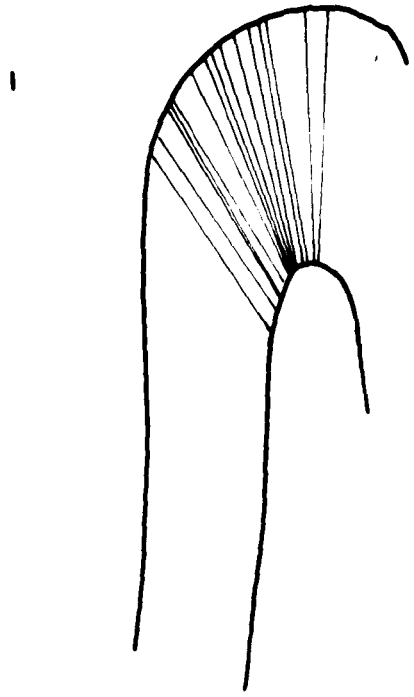
Fold 20



Fold 30



Fold 7



Fold 11

Fig.3 F_2 fold profiles with dip isogons

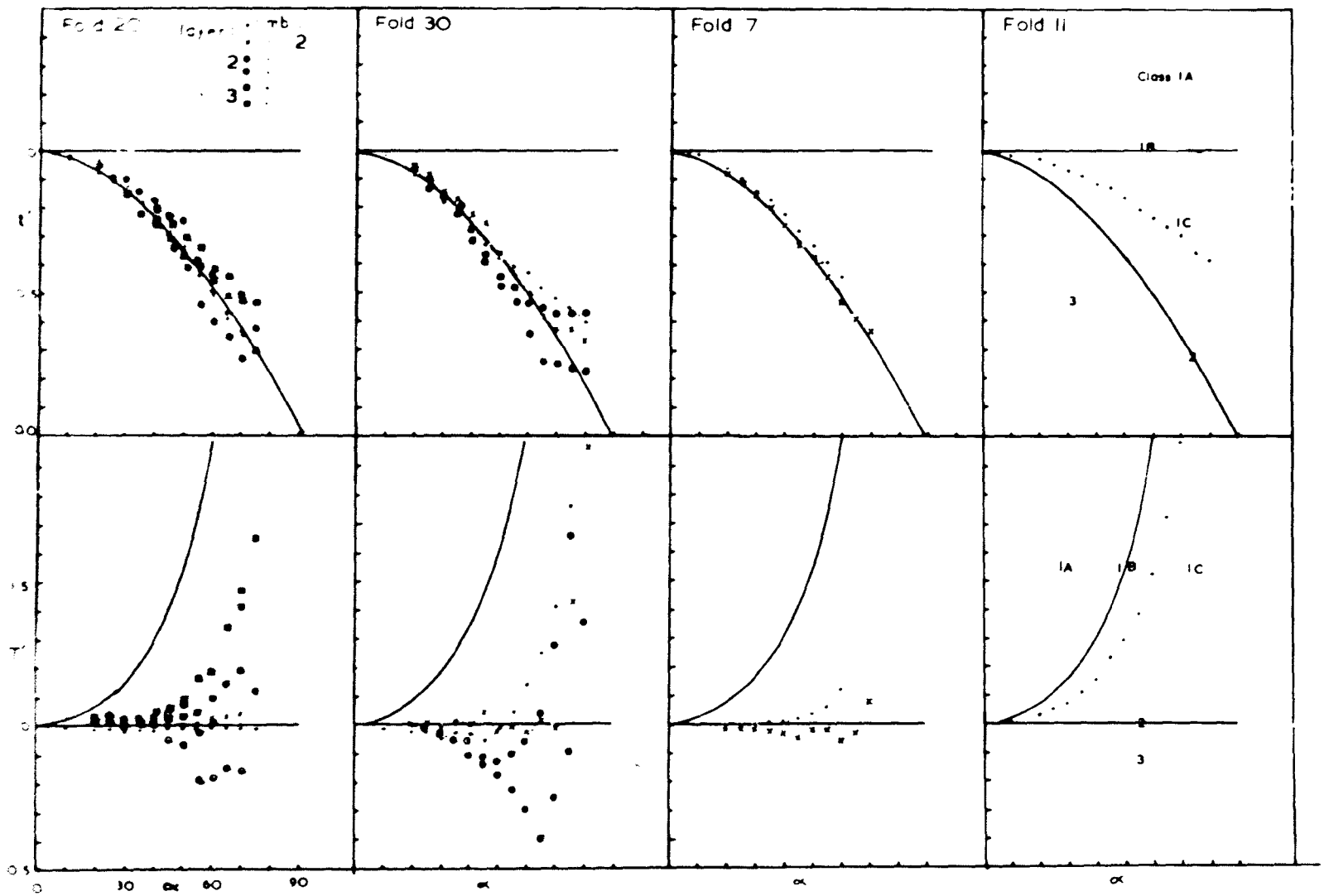


Fig.4 plots of t' and T' against α for F_2 folds

(see Ramsay, 1967, p 359-72)

The variation in style, both within and between F_2 structures, provides some clues as to both the original nature of the F_2 deformation and that of the succeeding phase. The change from classes 2 or 3 at the hinge to class 1C on the limbs in folds from migmatitic gneiss (Figs. 3 and 4) is similar to that shown by Ramsay (1967, Fig. 7-27) for a class 1A fold modified by homogeneous strain. The changes in style within individual folds could be accounted for in part by the effects of a homogeneous strain on an original alternation of various classes of fold in different layers in the F_2 fold forms (see Ramsay 1967, p. 415). The regular class 1C of fold 11 suggests the modification of a parallel fold in calc-silicate rock by homogeneous strain. The presence of multiple-hinged folds with complementary isoclinal folds, folds with highly attenuated limbs, and flattened ptygmatic folds in some granofels bands, illustrates the selective nature of the secondary deformation. The consistent presence of class 2 folds in little deformed metabasite bodies may reflect the original nature of the F_2 folds in such lithologies. Ptygmas and viscosity contrast structures affecting granitic material in unfolded granofelses indicate that during D_2, quartzo-felspathic rocks behaved more competantly than intermediate rocks which deformed without the formation of folds or fabric. If, as suggested, F_2 folds in migmatitic gneisses were compounded of alternating styles, this competance of acid material would be consistent with the gneisses' inability to flow plastically during D_2.

D_3: Formation of foliation S_3.

When axial plunges of F_2 folds from the gneisses in the core of the Port na Luing - Totronald major structure are plotted stereographically they lie on a great circle, whose pole is in the vicinity of the maximum distribution of poles to foliation planes and F_2 axial surfaces (Figs. 5a and b). This would suggest that the linear fabric consisting of F_2 axes was deformed in the plane of the foliation by a homogeneous straining of the foliation. This is consistent with the observed deformation of individual isoclinal folds, and the occurrence of relatively strongly foliated bands in the migmatitic gneisses. Variation in axial plunge within deformed F_2 isoclinal folds is well shown by Plate 4c of elongate eye-like refolded structures occurring along the axial surfaces of isoclines. They are interference patterns of Ramsay's type 1 (1967, p. 552), axial traces of

The variation in style, both within and between F_2 structures, provides some clues as to both the original nature of the F_2 deformation and that of the succeeding phase. The change from classes 2 or 3 at the hinge to class 1C on the limbs in folds from migmatitic gneiss (Figs. 3 and 4) is similar to that shown by Ramsay (1967, Fig. 7-27) for a class 1A fold modified by homogeneous strain. The changes in style within individual folds could be accounted for in part by the effects of a homogeneous strain on an original alternation of various classes of fold in different layers in the F_2 fold forms (see Ramsay 1967, p. 415). The regular class 1C of fold 11 suggests the modification of a parallel fold in calc-silicate rock by homogeneous strain. The presence of multiple-hinged folds with complementary isoclinal folds, folds with highly attenuated limbs, and flattened ptygmatic folds in some granofels bands, illustrates the selective nature of the secondary deformation. The consistent presence of class 2 folds in little deformed metabasite bodies may reflect the original nature of the F_2 folds in such lithologies. Ptygmas and viscosity contrast structures affecting granitic material in unfolded granofelses indicate that during D_2, quartzo-felspathic rocks behaved more competently than intermediate rocks which deformed without the formation of folds or fabric. If, as suggested, F_2 folds in migmatitic gneisses were compounded of alternating styles, this competance of acid material would be consistent with the gneisses' inability to flow plastically during D_2.

D_3: Formation of foliation S_3.

When axial plunges of F_2 folds from the gneisses in the core of the Port na Luing - Totronald major structure are plotted stereographically they lie on a great circle, whose pole is in the vicinity of the maximum distribution of poles to foliation planes and F_2 axial surfaces (Figs. 5a and b). This would suggest that the linear fabric consisting of F_2 axes was deformed in the plane of the foliation by a homogeneous straining of the foliation. This is consistent with the observed deformation of individual isoclinal folds, and the occurrence of relatively strongly foliated bands in the migmatitic gneisses. Variation in axial plunge within deformed F_2 isoclinal folds is well shown by Plate 4c of elongate eye-like refolded structures occurring along the axial surfaces of isoclines. They are interference patterns of Ramsay's type 1 (1967, p. 552), axial traces of

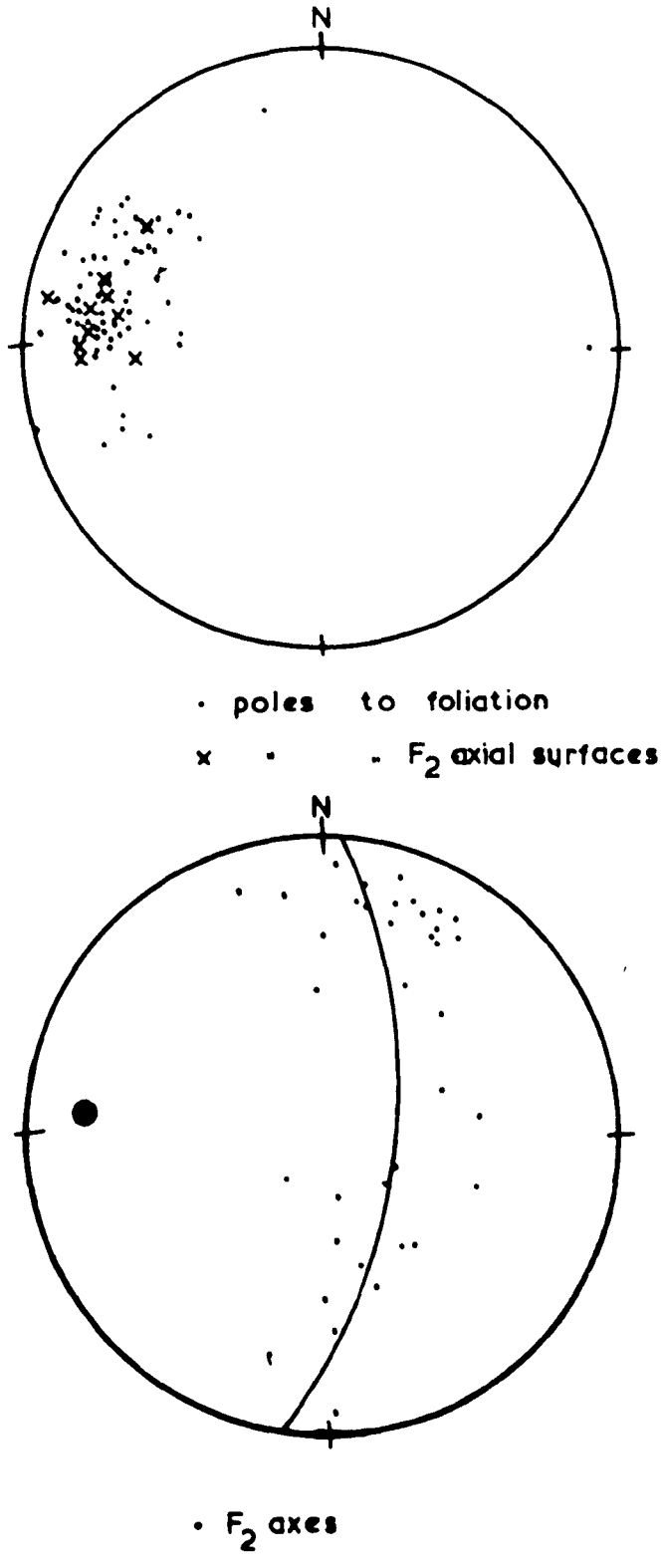


Fig.5: lower hemisphere equal area stereographic projections of F_2 axes, F_2 axial surfaces and S_1 foliation; Totronald, Coll.

superimposed folds being absent. Their form is consistent with variation in compressive strain along the axial surface of F_2 during D_3.

After the initial formation of the isoclinal folds, white quartzose pegmatites were intruded, as is evidenced by Plate 4d where such a body crosses the axial surface of an isolated F_2 fold. The pegmatite itself is folded tightly, apparently about an axial surface parallel to that of the isoclinal fold and the dominant foliation. Plate 5a shows a similar pegmatite cutting unfolded, strongly foliated migmatitic gneiss, and folded by an isoclinal fold whose axial surfaces are parallel to the foliation. Both figures strongly suggest large scale extension in the plane of the foliation during D_3. Because the limbs of the folds in the quartzose pegmatites are attenuated to varying degrees, it is not possible to calculate the exact amount of extension which occurred, but it is estimated to be of the order of 300 per cent in Plate 4d and 500 per cent in Plate 5a. The attitude of the pegmatites at the hinges of both folds suggests that they may have been emplaced perpendicular to the foliation. These bodies are rarely preserved in a discordant mode, generally being discontinuous, concordant streaks in most occurrences (see Map 1), possibly as a result of much greater extension.

The concordant granitic veins in the granofels as mentioned on p. 26, sometimes occur in intrafolial, deformed ptygmas. In a few localities they can be seen to be boudinaged in a sigmoidal fashion (Plate 5b).

Although a foliation was present in the gneisses, metasediments and metabasite bodies before the extreme straining in this phase of deformation, as seen in the isoclinal folds, it was further accentuated by this later phase. Evidence for this is seen in several features:

i) The deformation of isoclinal folds and their frequent separation by belts of well foliated gneiss.

ii) The presence of boudinage and pinch and swell structures in mafic and metabasite bands in well foliated gneiss (Plates 5c, 5d). These figures show flexures of foliation in migmatitic gneiss into cicatrices between individual boudins, and pegmatoid secretions developed in the cicatrices and as thin veins perpendicular to metabasite margins. In isoclinally folded metabasites, boudins are never seen to be folded about isoclinal fold axes. Boudinage on a major scale is illustrated in the gneisses west of Crossapol Bay and near Arinthluic on Map 1.

iii) In nebulitic gneisses, mafic and metabasite bands are not seen,

PLATE 5

Plate 5

a. Isoclinally folded quartzose pegmatite, cutting unfolded migmatitic gneiss, deformed during D_3: 194598, Coll.

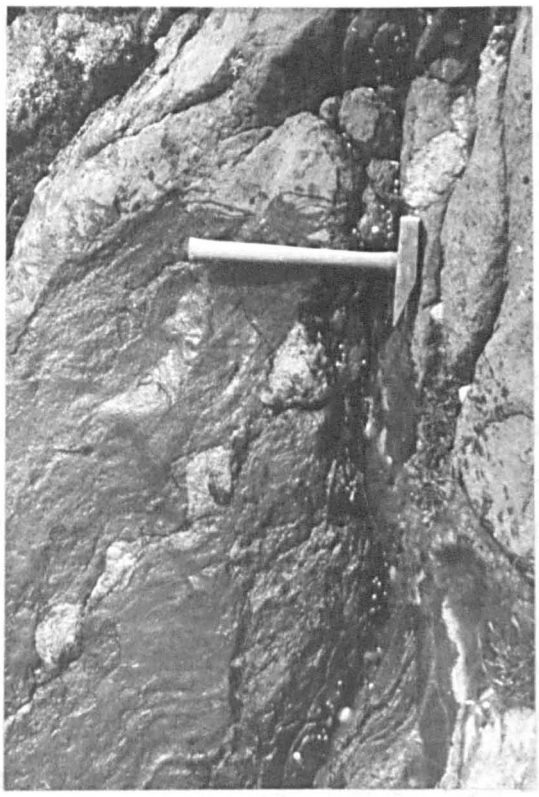
b. Sigmoidal boudins of concordant granite in granofels: Port na Luing, Coll.

c. Boudinaged metabasite, note flexure of foliation into cicatrices: Eilean Bhoramuill, Coll.

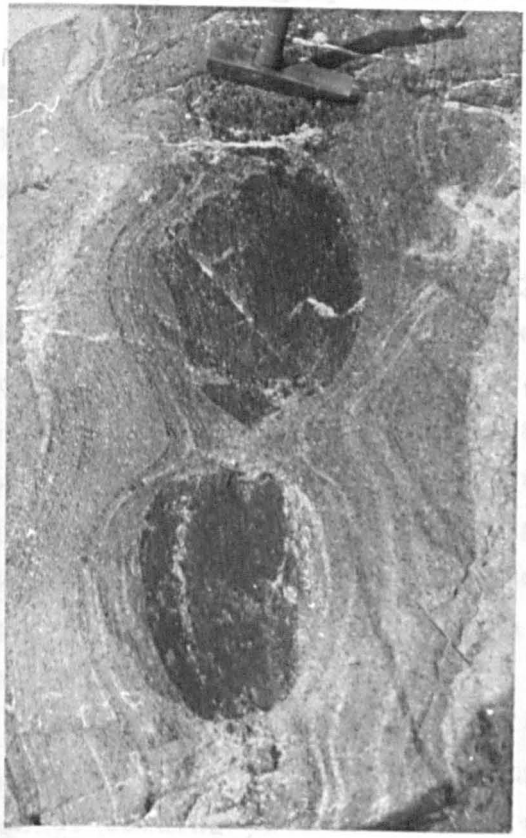
d. Boudinaged mafic band in migmatitic gneiss: 122520, Coll.

Plate 5.

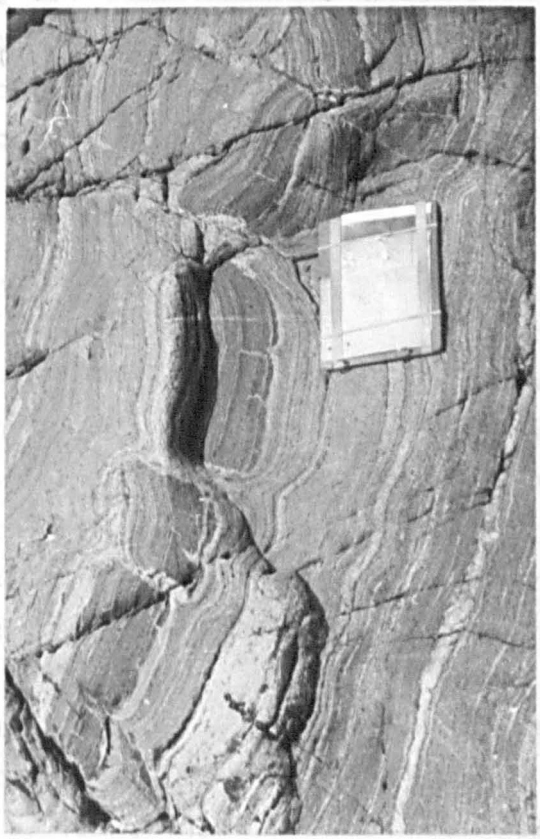
a



b



c



d

being represented by isolated mafic enclaves and flattened mafic to ultramafic schlieren which contain relics of isoclinal folds and rotated foliation (Plate 6a). In certain cases individual schlieren are seen to have been boudinaged with much associated pegmatoid material, and a progressive series of such structures seems to have resulted in the production of hornblendite pods (Plate 6b).

iv) Concordant granitic veins and bands in granofelses and metabasites do not show a foliation when folded by F_2 folds. Where they occur on the limbs of the major F_2 fold between Port na Luing and Totronald, the thicker bands, as between 157536 and 156525 , and at 125520 and 163570, in granofels, are unfolded but possess a marked planar fabric. A similar fabric is seen in many granitic veins in a metabasite body near Sorisdale.

v) Numerous bands of very well foliated gneiss are prominent within the more massive migmatites, exemplified by the narrow central band within the Hogh Bay - Breachacha metasedimentary belt, which has a phacoidal foliation and quartzo-felspathic "eyes" (Plate 6c). Such bands are devoid of F_1 intrafolial or F_2 isoclinal folds, which might be assumed to have been disrupted by strong D_3 deformation. They are characterised by the presence of a linear fabric L_3, defined by quartz rods, which plunges parallel to the foliation dip. A similar lineation is present in calc-silicate rocks and quartzites at 163571, and in what is interpreted as a highly deformed and boudinaged quartzose pegmatite in the Clabhach - Gorton metasediment belt.

It is suggested that the major foliation trends, on Coll at least, are the product of post-D_3 fold phases' effects on a regular penetrative foliation developed, at least in part, during this phase. The fact that those structures indicative of the greatest deformation occur in nebulitic gneisses would seem to suggest that this rock type was largely formed at this time and under the influence of strong deforming forces. These acted on a material more easily deformed than the bulk of the gneisses on the islands. Further discussion on the origin of the nebulitic gneisses is contained in Chapters 3 and 4.

As described, boudinage of this generation is well developed in metabasite bands. However, a limiting thickness of metabasite above which regular boudins have not formed seems to prevail. In certain thick concordant metabasites an agmatitic structure has been developed, in which relics of isoclinal folds are seen in the blocks of mafic rocks separated

PLATE 6

Plate 6

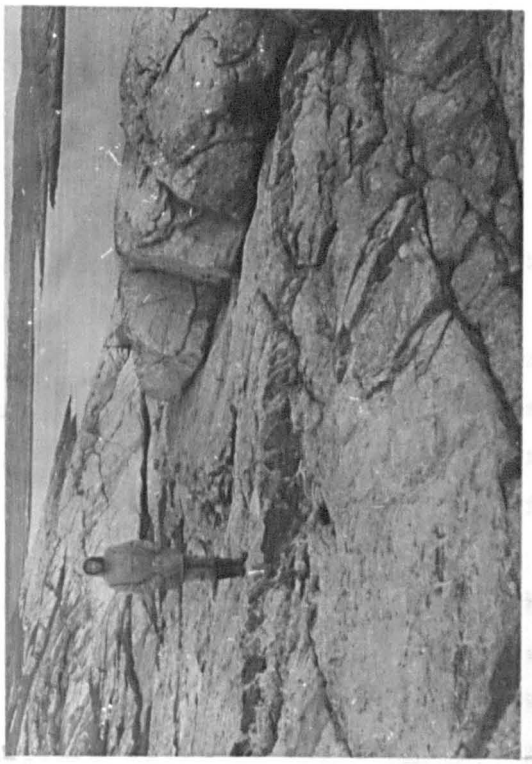
a. Rotated metabasite enclave in nebulitic gneiss: 177573, Coll.

b. Boudinaged hornblendic schlieren in nebulitic gneiss: 197534, Coll.

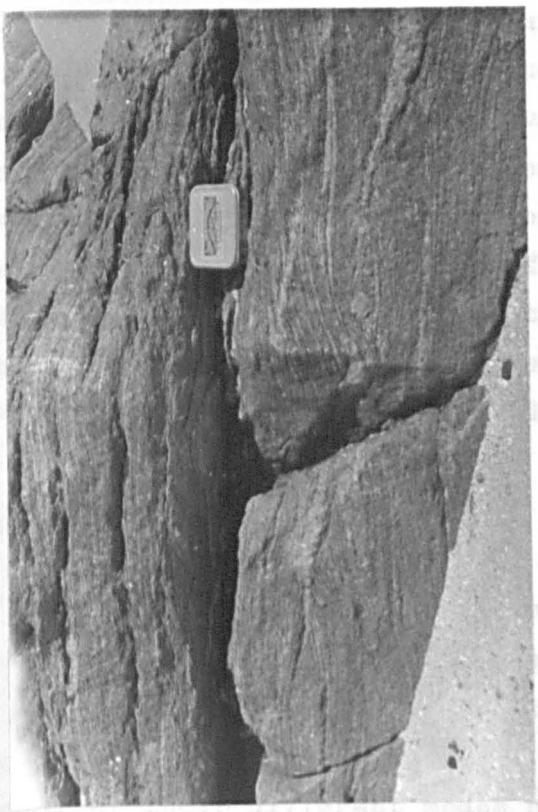
c. Phacoidal foliation and augen in deformed gneiss: Breachacha, Coll.

d. Type 1 interference structures, F_4 on F_2: 157563, Coll.

a



b



c

d

by white pegmatoid veins. The field appearance of agmatites and their origins are discussed further in Chapter 3. They generally occur in concordant masses within the migmatitic gneisses, and are particularly well developed within or close to nebulitic gneiss bands. Some homogeneous metabasite bands grade into agmatites along the strike. Thin agmatite bodies have a weak planar fabric in the felsic parts which can be seen to be folded by small scale F_4 asymmetrical folds and agmatites lie in the flat and steep limbs of large F_5 monoclinal folds. However some, which consist largely of coarse hornblendic pegmatite and small jumbled bodies of mafic material, are discordant to the foliation. Some of the felsic material can be seen to emerge from agmatites and cut across asymmetrical fold axial surfaces, and often lies along such surfaces. Plate 10a shows a concordant metabasite band which is folded by F_4 asymmetrical folds with intimately associated diffuse quartzo-felspathic material. The margins of the band are basified in contact with this material and the introduction of pegmatite parallel to axial surfaces appears to be at the stage of breaking up the body prior to the formation of an agmatite. On the basis of this conflicting evidence it is suggested that agmatisation proceeded over a long interval, but being initiated during D_3.

During this widespread flattening of the foliation it has been shown that migmatitic gneiss was very variable in its competancy reaching its most fluid state in nebulitic gneiss. Competant metabasites in close proximity to such rocks, would tend to disintegrate in a relatively brittle fashion, allowing silicate-rich fluids from the nebulitic gneisses to penetrate them. Seemingly, the activity of such fluids, and deformation which may be subdivided into several phases, continue for some time, agmatites being formed and reworked during this period.

D_4: Asymmetrical Folding, F_4.

Folds of this generation are widespread but inhomogeneously distributed on Coll, being most common in the migmatitic gneisses. They fold both S_1 and S_3. Certain types are to be found in nebulitic gneisses. In the migmatitic gneisses, their style and attitude is highly variable both as a result of their different modes of formation and subsequent deformation. In general they are only found in a size ranging from 10 m wavelength to a few centimetres. However, because of their relationships to earlier and

later major fold axial traces, certain major folds have been ascribed to this generation. In general they have an overturned asymmetrical form with shallow and steep limbs. They refold F_2 folds and their axial surfaces are rotated by F_5 monoclines and F_6 open folds.

In the complex refolded structure between Brock and Fhadamuill on Tiree, a major fold, whose axial trace has been folded by an E-W trending F_6 open fold, itself folds the axial trace of an earlier F_2 isoclinal structure, and it can be provisionally correlated with F_4. It has sparse parasitic folds identical to the F_4 asymmetrical folds on Coll. Similarly, a large fold, whose axial trace runs N-S from Traigh Halum to Eilean Bhoramuill and whose eastern limb dips steeply eastwards and western limb dips steeply to the south, folds structures such as boudinaged metabasite bands and small F_2 isoclinal fold axes. Parasitic folds around the hinge indicate that the major structure plunges steeply to the south, south of the axial trace of the major F_2 axial trace shown on Map 2. On the northerly dipping limb of the F_2 isocline, small asymmetrical folds plunge steeply to the north, indicating that they were superimposed on an earlier, isoclinally folded surface. The presence of thick, strongly boudinaged metabasites, and the absence of folding in the gneisses of the easterly dipping limb, suggests that this limb is relatively more attenuated than the southerly dipping limb. The latter is strongly contorted by small F_4 parasitic folds.

Direct evidence for refolding of F_2 folds by F_4 asymmetrical folds is limited by the general absence of well preserved F_2's and the sporadic distribution of asymmetrical folds. F_4 folds are most prominent where the foliation is strong, and isoclinal folds are absent because of the strong D_3 deformation. Structures produced by the interference of F_2 isoclinal folds and F_4 asymmetrical folds are well developed at 160562 where a belt of asymmetrical folds of varying size intersects the axial surface of the major F_2 fold running from Port na Luing to Totronald (Map 2). Plate 6d shows several recognisable isoclinal and asymmetrical folds and also numerous interference structures, mainly of Ramsay's type 1, but some showing affinities to type 2 (see Ramsay, 1967 Fig. 10-13, p. 531). Plate 7a shows an interference pattern of type 2 produced by the superimposition of an asymmetrical S-fold on an isoclinal Z-fold. Type 3 interference patterns are seen in the hinge zone of the major F_4 fold, north of Eilean Bhoramuill (Plate 7b). Plate 7c shows a well developed type 2 structure from 218622. Plate 7d shows a type 3 structure from calc-silicate rocks at Clabhach.

PLATE 7

Plate 7

a. Type 2 interference structure, F_4 on F_2: 157563, Coll.

b. Type 3 interference structure, F_4 on F_2: 120519, Coll.

c. Type 2 interference structure, F_4 on F_2: 218620, Coll.

d. Type 3 interference structure, F_4 on F_2: Clabhach, Coll.

a

b

c

d

To the NW of Totronald a series of intermediate scale F_4 asymmetrical folds, whose axial traces tend SW-NE, fold the attenuated migmatitic gneiss of the eastern limb of the major F_2 isoclinal fold (Map 2). They have an amplitude between 1 and 10 m and a wavelength roughly twice as large. The interlimb angle is usually greater than 50° and less than 120°, and they may be termed open to close folds. They are frequently overturned and plunge at angles up to 30° towards the NE. Their axial surfaces are planar, and dip consistently at angles around 45° to E and SE. Variation in symmetry is common but they are mainly harmonic, periodic asymmetrical waves of S-type with steep short limbs (Plate 8a). They are defined by the foliation in the gneisses and have second and third order folds of S-, M- and Z type (Plate 8b). S-types are found in the shallow limbs and Z-types in the short steep limbs. They do not possess any penetrative axial planar structure. Although no marked attenuation is present, flattening is relatively greater in the gently dipping limbs of first, second, and third order folds. The fold shown in Plate 8b has dip isogon orientations and t' and T' versus α plots strongly indicative of its close approach to Ramsay's class 2.

F_4 folds of a similar but more disharmonic nature are strongly developed SE of the road between Gallanach and Toraston in strongly foliated migmatitic gneisses with large proportions of concordant granitic veins (Plate 8c). The fold shown has a rough axial planar structure defined by the attenuated flat limbs of minor folds in the steep limb. Ramsay and Graham (1970) suggest that such small shear-like folds are analogues to, or represent , the axial plane cleavage seen in lower grade rocks. Where the foliation dips at low or moderate angles to the E and SE, these folds are consistently S-shaped plunging NE, with steep NW dipping short limbs and attenuation in the gentle SE dipping limbs. Their axial surfaces dip at a greater angle than their shallow limbs, but in the same direction. In a few places in this area steep W to NW dipping belts are found, in which F_4 S-shaped folds are seen with steeply dipping attenuated limbs and sub-horizontal middle limbs. Their axial surfaces dip at a lower angle than the foliation, towards the NW (Plate 8d). Evidently these folds have been rotated by a later fold phase. A similar situation is met with in the steeply dipping gneisses on the coast from A'Chroic to Rubha Mor, where F_4 folds of intermediate size all have S-symmetry, steep, attenuated limbs, flat short limbs and gently NW dipping axial surfaces (Plate 9a).

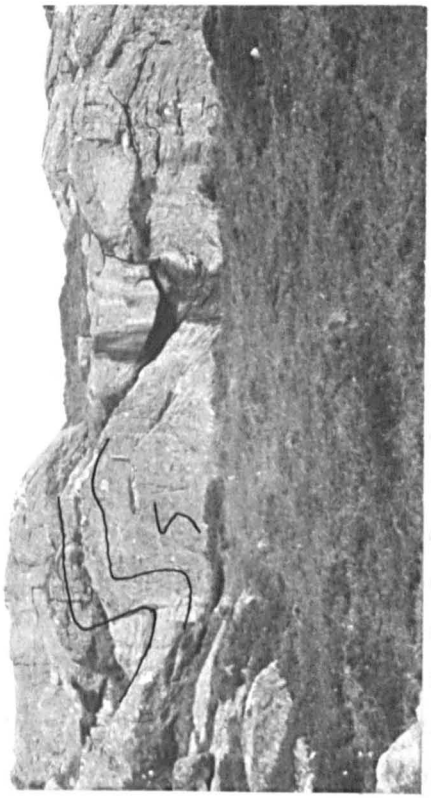
Plate 8

a. S-shaped asymmetrical fold: 162566, Coll.

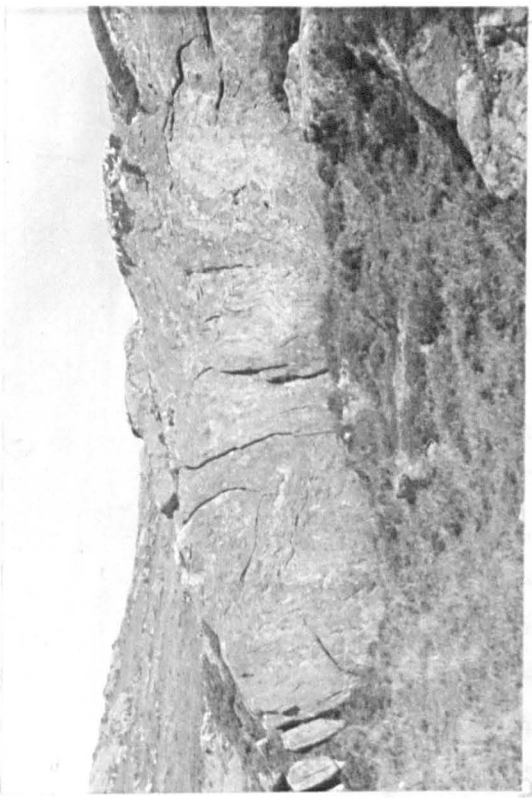
b. Z-shaped F_4 fold with minor F_4 folds: 162568, Coll.

c. Disharmonic F_4 with axial plane structures: 220609, Coll.

d. Rotated F_4 folds and F_4 ptygmatic folds in remobilised concordant granitic veins: 220613, Coll.

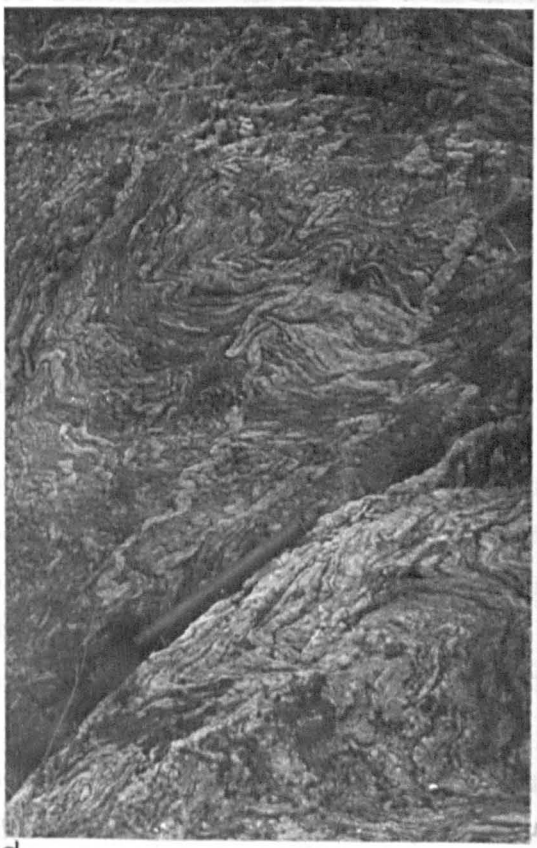


a



b

c



d

The greatest proportion of F_4 folds are of a small size, ranging up to 50 cm in amplitude, but generally only 10 cm. They are not necessarily associated with larger structures, often being found in isolation in other featureless gneisses. In fairly homogeneous migmatitic gneisses they frequently approach a similar style very closely, maintaining a harmonic form over relatively long distances along their axial traces (Plate 9b). Where the gneisses are very heterogeneous because of numerous quartzo-felspathic veins a ptygmatic style is prominent. Plate 8d shows both concordant and remobilised, discordant veins folded into ptygmas during this phase of deformation. In the migmatitic gneisses one of the most prominent features of small F_4 folds is the association of planar granitic bodies aligned along their axial surfaces (Plates 9a, c) and sometimes lying in the short steep limbs. They may be sharp-sided or diffuse and unfoliated. Their marked alignment along the hinge zones suggests that they were controlled by the folding or the processes causing their formation encouraged folding. Similar relationships have been observed by the author in migmatite terranes in the Lewisian elsewhere and in the Precambrian gneisses of the Angmagssalik area of East Greenland. Bowes and Hopgood illustrate granitic segregations from Mingulay (1969, Figs. 5, 15 and Plate 3 Fig. 3) associated with their F_2 phase of folding, which is apparently similar in style to that on Coll. Bowes (1966) has illustrated such relationships in the Mainland Lewisian. The dip of these granite veins is parallel to that of the axial surfaces of F_4 folds, and they have been rotated by later folds.

Metabasites and mafic layers in the gneiss are harmonically folded by intermediate sized asymmetrical F_4 folds (Plate 9d) frequently being agmatised in the steep limb, with much irregular discordant granitic material. In the steep limbs of some intermediate sized folds, metabasites can be seen to have developed lobate margins, rounded tongues of mafic rock being separated by narrow infolds of gneiss (Plate 10a). This could indicate compressive strain in the steep limbs producing such structures at interfaces between media differing in viscosity. Granitic material is invariably associated with the infolds of gneiss, and the margins of metabasites are darker than their interiors.

F_4 folds are of similar or class 2 style in the migmatitic gneisses, being harmonic through relatively large distances along their axial surfaces. Ramsay (1967, p. 430) considered that such folds may be the result of modification of buckle folds by homogeneous strain, in contrast to the generally

PLATE 9

Plate 9

a. Rotated F_4 folds in steep limb of F_5 monocline: 237634, Coll.

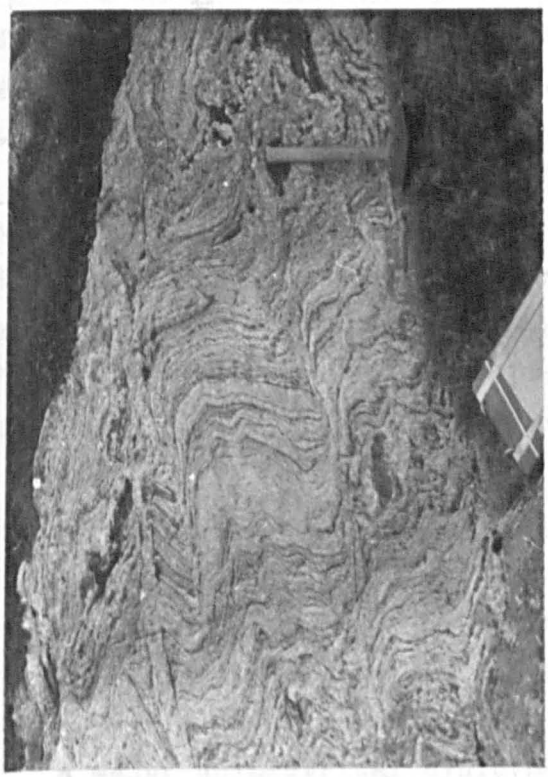
b. F_4 similar folds in migmatitic gneiss: 221612, Coll.

c. F_4 structures with planar axial surface granitic bodies: 163562, Coll.

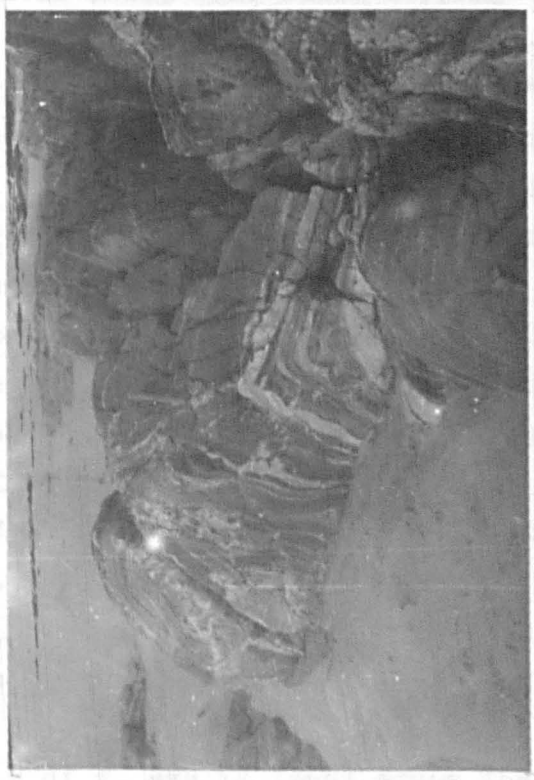
d. Harmonic F_4 fold in metabasite: 218620, Coll.

a

b



c



d

proposed mechanism of flow. The lack of evidence for modification of concentric buckle folds, and the mobile nature of the gneisses during D_4, evidenced by the presence of numerous diffuse granitic bodies, often intimately associated with folds of this generation, suggests that the folds are the product of flow under heterogeneous simple shear.

D_5: Monoclinal Folding, F_5.

Deformation in this phase formed broad monoclinal folds, consisting of local steep belts in gently dipping areas, and some upright folds with opposed dips on their limbs. F_4 axial surfaces are seen to have been rotated in some F_5 folds, and F_5 axial traces are folded about F_6 open fold axes. The most common F_5 structures are of an intermediate to large scale, steep belts being up to 200 m across. Some small folds of distinct style are assigned to this generation.

The most common F_5 structures are belts of gneiss dipping at high angles, sometimes vertically, in the otherwise gently dipping gneisses of that part of Coll east of a line from Cliad Bay to Arinagour. In general they take the form of open flexures with sub-horizontal axes, flat limbs dipping at angles from 5° to 45° SE, and steep limbs dipping in the same direction at high angles, sometimes vertically. Their axial surfaces dip at about 45° NW. The steep belts range from a few metres across to 200 m in the case of the structure NW of Sorisdale (Map 2). The monoclines around Cranaig have an axial plunge of about 20° NE, and are overturned in places. Parasitic folds of this generation can be seen in a steep belt on the coast at Sorisdale, folding metabasites, granofelses, and associated gneisses (Plate 10b). Fig. 6 shows their relationships to F_2 folds whose axial traces parallel the foliation trend. At the head of Bagh na Coille they are seen to refold an intermediate scale F_2 isoclinal fold, forming a complex interference structure of Ramsay's type 1. Varied symmetries and axial plunges are seen, ranging from S-shaped steeply inclined asymmetrical folds in the north, to upright open folds around Bagh na Coille. In the rock types present, the folds approximate concentric buckle folds with strongly convergent dip isogons.

A major asymmetrical synform probably of F_5 age is found along the SE coast of Coll where it has a steep to vertical NW dipping limb and a gentle limb dipping SE. It is well displayed in the cliffs of Meall Eatharna,

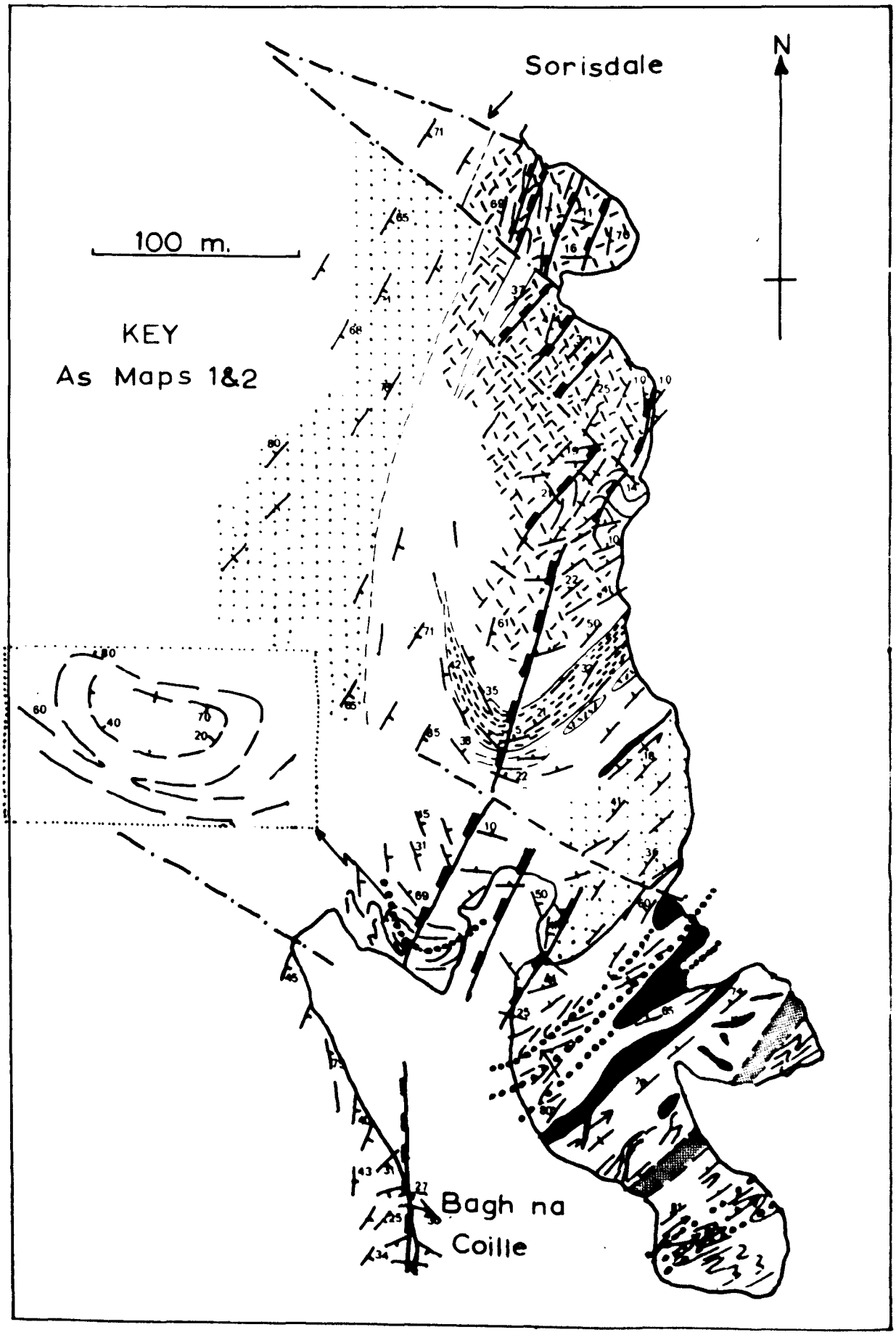


Fig.6: structural map of Sorisdale, Coll.

where its axial surface is exposed dipping at about 50° SE. An F_5 asymmetrical fold is represented on the NW coast of Coll by its steep to vertical NW limb, dipping generally towards the NW. Its hinge zone is largely obscured by a strongly developed later shear belt, but is preserved N of Cornaigbeag where the axis plunges at a low angle to NNE, and its shallow limb dips at moderate angles to SE. As has been mentioned, page 33, F_4 folds are present in the steep limb here where they now have axial surfaces dipping NW. In the exposures between Gallanach and Toraston, the trend of F_4 axial planes is seen to be curved, and numerous, narrow, discontinuous NW dipping belts contain rotated S-shaped F_4 folds (Fig. 7). The refolding of the earlier fold phase is demonstrated by Fig. 7 where F_4 axial traces have been rotated about a gently plunging axis situated in the NE sector of the stereogram. The F_5 folding was evidently co-axial with the F_4 folds whose axial plunges show no consistent variation (Fig.

In general, no appreciable attenuation of the foliation is associated with F_5 folds. However, in the steep belt NW of Sorisdale, certain transposition structures are seen (Plate 10c, 10d). Possibly pre-existing folds now have a strange style, and are separated by parallel bands of attenuated gneiss, representing deformed fold limbs, and defining a coarse planar foliation. All stages from fairly open buckles to tight flattened folds are seen along different axial surfaces (Plate 10d). Because of the discontinuous nature of the foliation and the presence of concordant granitic veins, no measurements were possible but approximate dip isogons vary widely in their orientation pattern, at different levels in the fold.

In the nebulitic gneisses of the Ben Hogh area, numerous small monoclines, usually only a few centimetres across, are seen to fold the gneisses. Their steep middle limbs are markedly attenuated and frequently occupied by diffuse granitic secretions (Plate 11a). In a very few cases they refold F_4 folds.

The general attitude of foliation in the islands, steep to vertical in Tiree and western Coll and gentle to moderate east of the line of the Hogh Bay - Breachacha belt, is suggested to be a result of this phase, modified by subsequent folding.

Biotite-pegmatites cut across structures in F_5 steep belts. They show a marked concentration along hinge zones of earlier folds (Map 2), and in some localities form pipe-like bodies along the axes of various generations of folds.

Plate 10

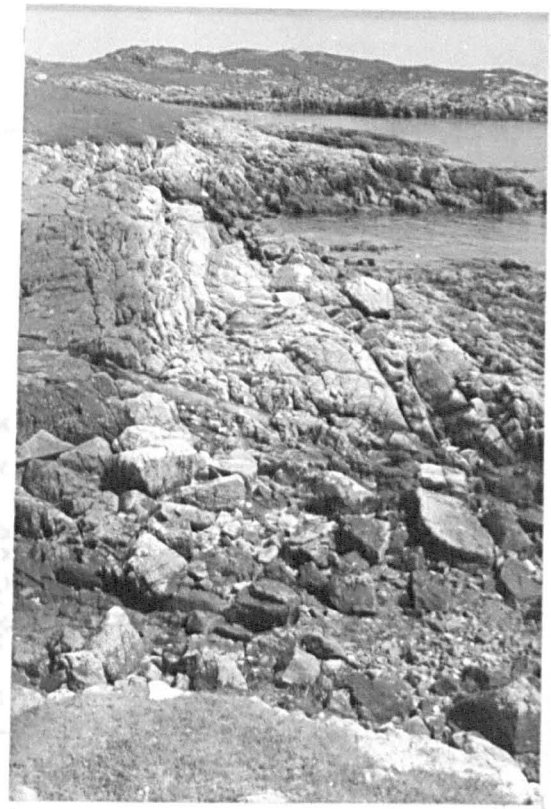
a. Lobate margins of metabasite, with associated small F_4 folds, granitic secretions and basified border: 222610, Coll.

b. Parasitic F_5 fold in steep limb of major F_5 monocline: Sorisdale, Coll.

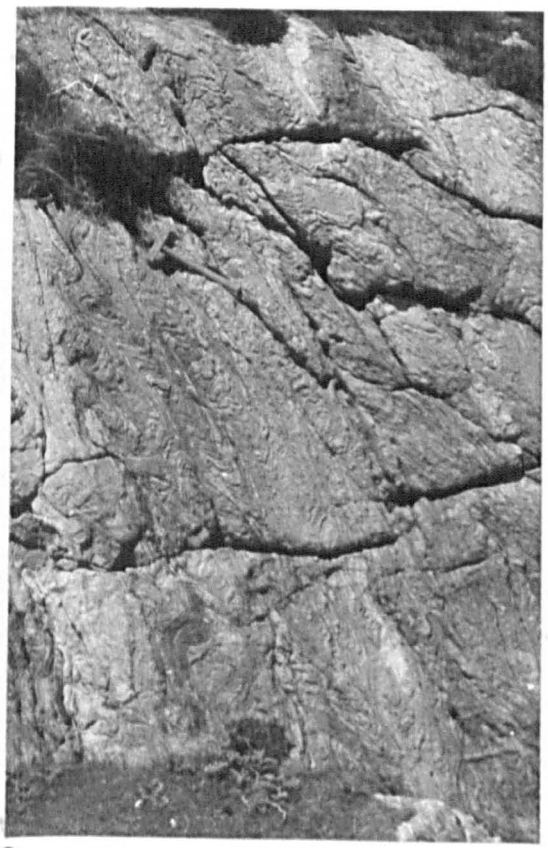
c. Transposition structures in steep limb of major F_5 monoclines: 272638, Coll.

d. Close-up of 10c, showing nature of transposed earlier folds.

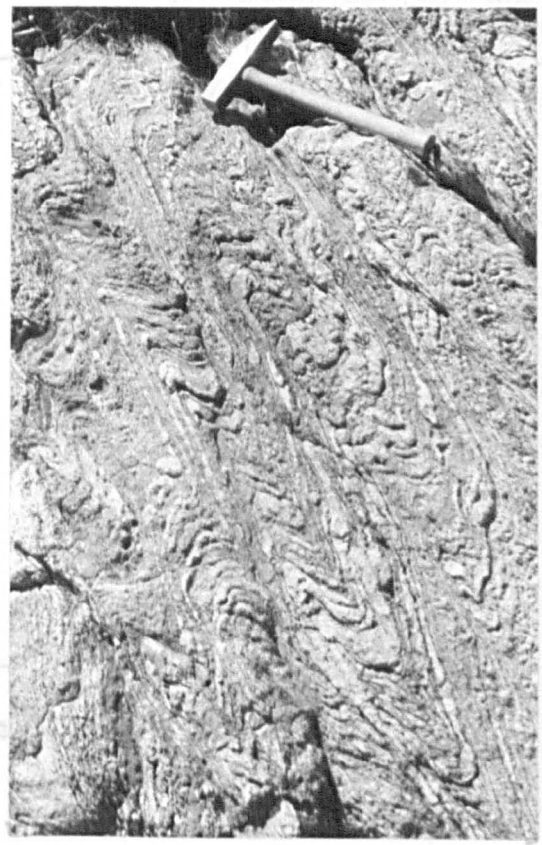
a



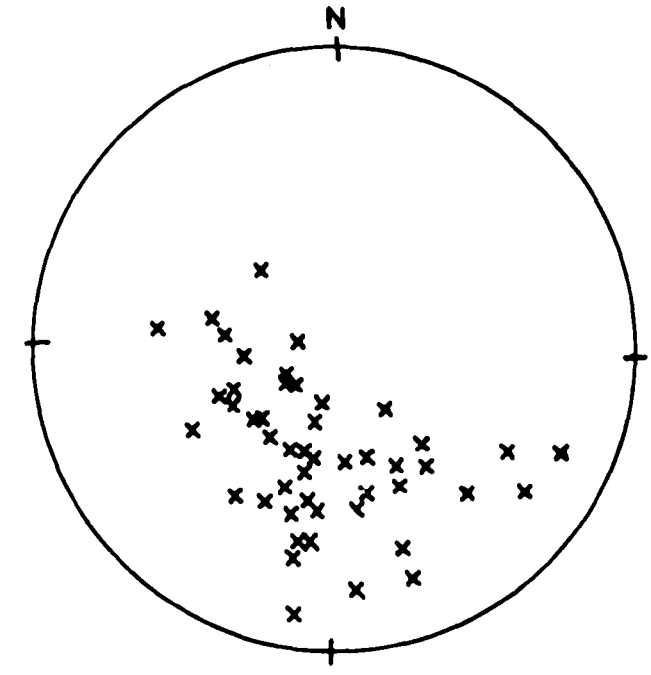
b



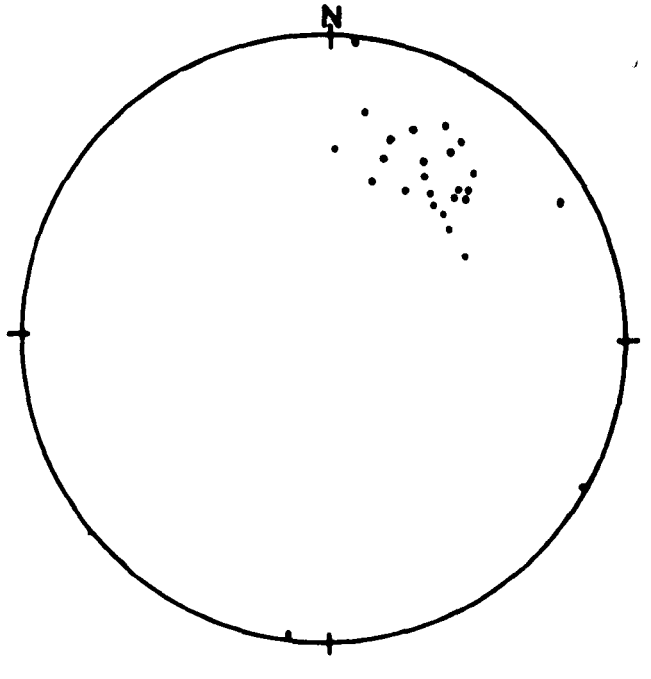
c



d



× poles to F_4 axial surfaces



• F_4 axes

Fig.7: stereographic projections of F_4 axes and axial surfaces, Gallanach to Torastan, Coll.

D_6: Formation of Open Folds, F_6 and Shear Belts, F_6 and S_6.

Folds of two distinct styles, open flexures and shears, are assigned to this deformation phase. They are never present in intimate relationships and no evidence for their age relative to one another is available. However, they have effects on all earlier structures and are themselves dislocated by late faults.

Broad upright, open, flexures of major size, trending roughly ENE to NNE and plunging at various angles are prominent on Maps 2 and 3. In places their axial traces bifurcate and they become multiple-hinged folds. They refold major F_2 folds between Brock and Fhadamuill on Tiree and at Totronald on Coll. Minor F_4 fold axial surfaces can be seen to be folded about F_6 axes, especially on Tiree and on the slopes of Ben Hogh on Coll (see Map 2). In the exposures between Cranaig and the NW coast of Coll, two F_5 monoclinal steep belts can be seen to curve around the hinge of an F_6 open fold trending NNE, the plunge of the open fold axis changing from about 15° NNW to vertical in places, as the monocline is crossed (see Map 2).

Interference structures caused by F_6 open folds refolding earlier generations are localised. Plate 11b shows a dome and basin effect caused by open folds' effects on box folds of obscure age in calc-silicate rocks at Clabhach. Plate 11c shows similar features resulting from superimposition of F_6 open folds on small F_5 monoclines at 188598.

In the gneisses east of the Clabhach - Gorton line, belts of sheared chlorite-epidote-gneiss are common on all scales from a few centimetres to several hundred metres. They may take the form of steep SE dipping, bifurcating belts possibly controlled in part by F_5 steep limbs, as around Sorisdale. A large synform attributed to the F_5 phase, running from Meall nan Uan to Meall na h-Iolaire can be seen to pass into a sheared gneiss belt, its axial plunge changing from 20° SW through horizontal to vertical as its axial trace is followed from SW to the sheared belt. A possible continuation S of Bagh na Coille plunges at 26° SW. Much of the NE tip of Coll is made up of sheared gneiss containing lenses of less deformed, often rotated, pre-existing rocks. A prominent belt parallels the trend of an F_5 axial trace from Traigh Bound to Toraston.

Many belts of sheared gneiss are in the middle limbs of asymmetrical fold pairs whose axial traces trend parallel to the open folds. Inland, from Eilean nam Muc to Meall na h-Iolaire, three prominent F_6 fold pairs,

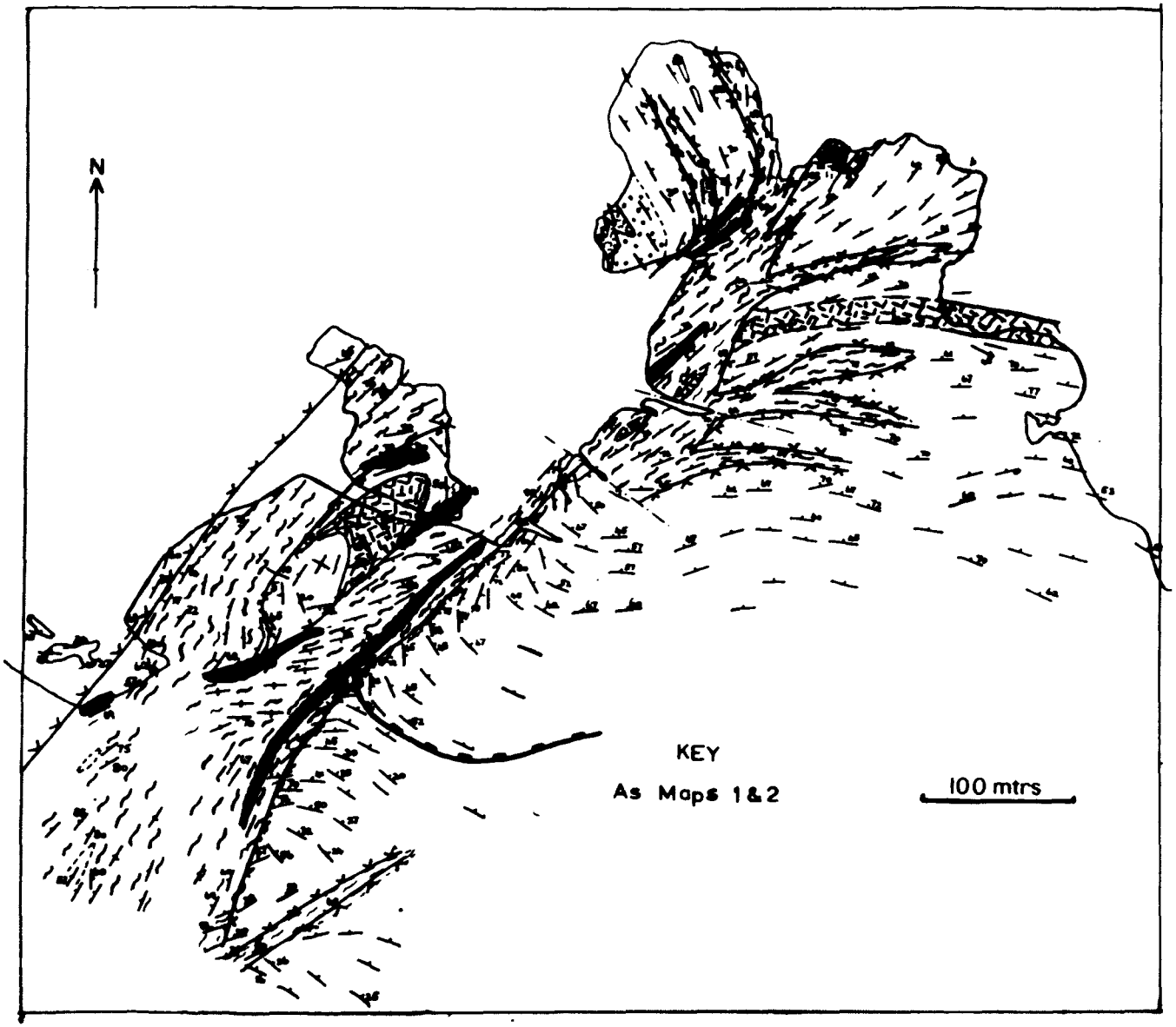


Fig.6: structural map of F_6 shear belt at Gallanach, Coll.

PLATE 11

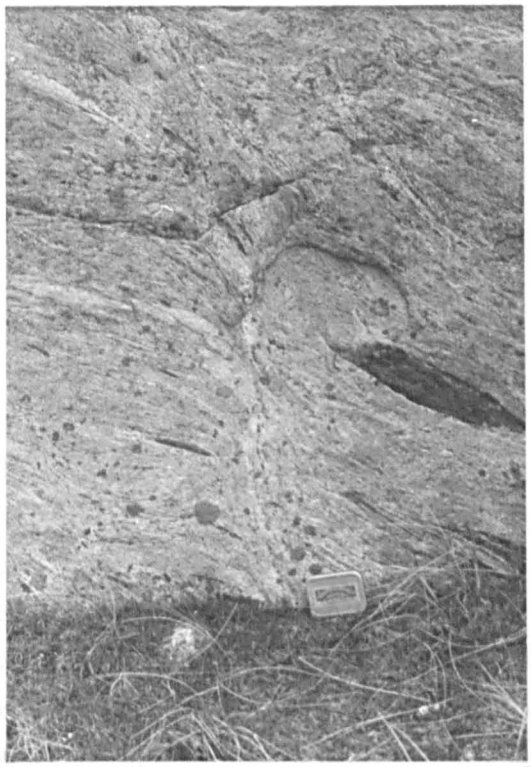
Plate 11

a. Minor F_5 monoclinal structure in nebulitic gneiss: 177573, Coll.

b. Dome and basin interference pattern in calc-silicate rocks: Clabhach, Coll.

c. Ditto in migmatitic gneisses, F_6 open folds on F_5 monoclines: 188597, Coll.

d. F_7 structure associated with flinty crush rock zone: 262615, Coll.



a

b

c

d

Probably the shear zones were formed by a combination of heterogeneous simple shear and homogeneous strain (Ramsay 1967, p. 423), when the rocks were in a relatively plastic state due to retrogression in certain areas under greenschist facies conditions where hydrous fluids were active.

D_7: Brittle Structures.

Black flinty crush rock is found in the steep limbs of monoclines and shear belts. It is frequently associated with contemporaneous folds (Plate 11d) often of chevron type. In appearance, this rock is aphanitic and occurs in irregular anastomosing veinlets including angular fragments of country rock. An intrusive habit is seen, thin veins penetrating into undisturbed gneiss with sharp boundaries and no displacement of foliation across them. Similar rocks have been described from the Meall Deise region by Bhattacharjee (1965) and Barra (Bowes and Hopgood 1969). Black flinty crush rocks are very prominent E of the Outer Isles Thrust in the so called "mashed gneiss".

Numerous vertical faults are apparent on Coll, being markedly concentrated along the SE coast from Sorisdale to Meall Eatharna. They form marked topographical features. Two main trends are seen, one set trending NNW, the other WNW. They seem to be wrench faults, the NNW ones having a dextral sense and the WNW a sinistral displacement. Invariably, the fault planes have associated green epidotic crush rock, and the gneiss has been turned pink in their proximity. Folds of all generation are displaced by the faults, and planar epidotic crush veins cut across flinty crush rock belts. In many cases, the foliation in the gneisses is rotated into parallelism with the fault planes over short distances, this effect being most marked in sheared gneiss around Meall na h'Iolaire (see Map 2). A series of thrust planes marked by epidotic crush rock carry unsheared gneiss over sheared gneiss at Beart nan Fhir.

Discussion.

It has been shown that the rocks of Coll, and, to a lesser extent, Tiree have evolved their structural form through a long, complex, history of deformation. Certain structures existed prior to the emplacement of intermediate, basic and ultrabasic rocks, which possibly represent a long

and complex structural history now undecipherable. The present form is the result of post-intrusion deformation, only a few square kilometres on Tiree maintaining the discordant relationships of basic and intermediate rocks. Several inferences can be drawn regarding the progress of deformation, and the changes in the rheological state of the different lithologies.

1) Concordant metabasites and granofelses contain granulite facies mineral assemblages, some of the metabasites also possess a mineralogical banding, and concordant granitic veins in a few cases contain hypersthene. These facts would be consistent with the production of F_1 intrafolial folds and S_1, and the rotation of dykes into concordance either before or during granulite facies metamorphism. To produce such a strong foliation and dislocated folds, without the presence of boudinage, one must conclude that the complex deformed almost homogeneously, all lithologies having similar competency. As concordant granitic veins seem to have formed during or closely following this deformation, although they are themselves unfoliated, migrating silicate-rich fluids were associated with D_1.

2) Structures of several generations have been described from the various lithologies present which show the effects of contrasted competencies during the deformation phases.

Probably all rock types were deformed equally during D_1 indicating a similar rheological state in both felsic and mafic rocks.

The ptygmatic and lobate D_2 structures in interbanded granofelses and and white granites show that the felsic rock had a greater viscosity than the more mafic granofels. This inference is perhaps supported by the F_2 folds of highly variable geometry in felsic migmatites contrasted with class 2 F_2 folds in metabasites. The lower viscosity of mafic rocks relative to adjacent felsic rocks could be accounted for by the anhydrous nature of the complex and its elevated temperature either during or at the close of the granulite facies metamorphism.

Metabasites were strongly boudinaged during D_3 when their viscosity was greater than that of surrounding felsic gneisses. This situation may have arisen as a result of the viscosity of the migmatitic gneisses being lowered by incoming hydrous fluids, which would tend not to affect the metabasites to as great an extent. Nebulitic gneisses, which are very felsic, show the most marked effects of lowered competency and also the strongest development of granitic bodies, and it is likely that they were subject to large scale partial melting or solution during D_3. Agmatisation

seems to have been initiated in D_3 by tectonic fragmentation of competent metabasites in incompetent felsic gneisses and probably progressed under the influence of mobile fluids from the surrounding gneiss as is evidenced by basic margins and the presence of numerous rootless granitic veins. Agmatisation proceeded through D_4.

The contrast in competency continued during D_4 and D_5. The D_6 shearing affected all lithologies to a similar extent, probably as a result of mineralogical changes giving chlorite-rich assemblages.

Mylonite formation in D_7 shows the brittle nature of all lithologies at this stage. Localised fusion caused injection of black flinty crush rock in an intrusive mode.

3) The tentative structural sequence given in Table 1 is only applicable to small areas where relationships are clear. Ages of major folds can only be certain when parasitic folds can be definitely related to them, and when they affect or are affected by other major folds of known age. Different sizes of contemporaneous folds may have differing geometries because of the affect of scale on competency.

Possibly there is a certain amount of flexibility in fold phases. For instance the homogeneous straining of D_3 with a superimposed heterogeneous simple shear could have produced F_4 folds with flattened shallow limbs. F_4 and F_5 minor folds often have associated granitic veins in their hinge zones, and may have formed when the gneisses were in a similar state, close to each other in time.

It is pointless to attempt correlation of the Coll and Tiree structural sequence with other areas of Lewisian because of the isolation of the area. There are certain similarities with Bowes and Hopgood's (1969) scheme for Mingulay, the closest of the Outer Isles. As they do not present any geometrical data for their fold phases it is difficult to carry the comparison further than a passing point of interest.

The lack of post-granulite facies basic dykes precludes structural comparison with the classical orogenic subdivision of the Lewisian.

4) Lewisian rocks of Coll and Tiree are of great interest in that they contain numerous generations of granitic rocks, listed in Table 1, relative to the tectonic sequence. Of these, three groups may provide evidence of the absolute age of parts of the sequence.

The concordant granites, so typical of granofelses and metabasites are pre-granulite facies. As they are potassic they may be amenable to radio-

metric dating, although similar rocks have not been reported from elsewhere in the Lewisian.

Quartzose pegmatites can be precisely dated as being post-D_2, pre-D_3. Identical rocks have been seen in Barra where they cut pre-granulite facies metabasite dykes and are cut by later Lewisian basic dykes. The Barra material has provided Rb-Sr whole rock ages of 2450 myr (Moorbath personal communication).

Biotite-pegmatites, which were emplaced between D_4 and D_5 are identical in many respects to such rocks from many Lewisian areas. Lambert (personal communication) has obtained Rb-Sr whole rock ages of around 2200 myr from many of them in widespread locations.

TABLE 1

Phases and features	Type of Deformation	Granitic fluid Activity
Early foliation, lineation folds	?	Diffuse speckled bodies
I n t r u s i o n o f m e t a b a s i t e s		
D_1, intrafolial folds and foliation	Transposition of earlier structures by flow?	Concordant granite veins
D_2, isoclinal folds	Buckling in acid material, shear folds in basic	
		Quartzose pegmatites
D_3, foliation, lineation	Homogeneous strain	Agmatisation
D_4, asymmetrical folds	Heterogeneous simple shear	Axial surface bodies
D_5, monoclinal folds	Transposition of earlier structures	Axial surface bodies
		Biotite-pegmatites
D_6, open folds, shear belts	Buckling, ductile faulting	
D_7, faults, mylonite belts	Brittle fracturing	

CHAPTER 3

PETROGRAPHY

In this chapter each of the major groups of rocks occurring on Coll and Tiree is dealt with in turn, a separate section being devoted to each. Firstly the distribution of lithologies, and the nature and variation of the rocks in the field is described. This is followed by descriptions of the mineralogy and texture of the rocks in thin section, and each section is concluded by a brief discussion of petrogenetic hypotheses suggested by petrographic evidence. A discussion to the whole chapter gives a comparison of lithologies with other areas of Lewisian, and the physical conditions of the metamorphic events and their time relationships are discussed.

Quartzo-Felspathic Gneisses.

Hypersthene-gneiss.

Massive, light coloured, partly homogeneous orthopyroxene-bearing gneisses are to be found only on Tiree, between Balephuil Bay and Hynish Bay, noteably on Ben Hynish (see Map 3). They outcrop in a roughly lens-shaped area, surrounded by well foliated migmatites. Outcrops of hypersthene-gneiss are massive and rounded. The rocks display faint colour banding on a scale of centimetres caused by inhomogeneities in the distribution of mafic minerals (Plate 19a). Within each band the ferromagnesian minerals, including stubby brown orthopyroxenes, occur as elongate granular aggregates defining a rough lineation lying in the plane of the banding. This fabric is paralleled in places by quartz-rodding. Fresh surfaces of the gneiss are dark compared with the pale weathered faces, having a 'greasy" appearance. In places a local stronger foliation is seen in these gneisses, and they come to resemble the banded migmatites outside this outcrop.

Included in this group are orthopyroxenic migmatitic gneisses from elsewhere on Tiree, because of their mineralogical affinities to the granulite facies rather than their field appearance which is similar to that of the amphibolite facies migmatitic gneisses (Plate 19c).

The hypersthene-gneisses show much evidence of local activity of silicate-rich fluids, such as diffuse discordant quartzo-felspathic secretions

apparently derived from the host rock. On Ben Hynish they carry orthopyroxene and cut the massive hypersthene-gneisses, whilst further east, along Hynish Bay, they carry large garnets and hornblendes and cut well banded migmatitic hypersthene-gneiss. They are not seen as discrete bodies, but usually as irregular, unfoliated speckled areas grading into the gneiss. (Plate 19c).

In thin section interlobate granoblastic textures are common.

Quartz occurs as large lobate to amoeboid grains, rarely showing relict straight boundaries and triple junctions with interfacial angles of 120° similar to textures of equilibrium crystallisation in metals, applied to silicate mixtures by such authors as Spry (1969) and Maclean (1965). The grains are strongly undulose in extinction, bands and polygonal sub-grains extinguishing separately, indicative of the process of polygonisation. This is a result of straining of the quartz lattice and the production of dislocations in its structure which manifest themselves as irregularly distributed inhomogeneities in the lattice. The first step in polygonisation is for dislocations to accrete in planar arrays, acting as boundaries between relatively strain-free subgrains. (In the hypersthene-gneiss quartzes, the subgrains have gradational boundaries). Those subgrains having different orientations give the undulatory extinction observed. Spry (1969) has suggested that at elevated temperatures, dislocation movement and slip, and polygonisation processes become concomitant, the combined process being analogous to "hot-working" in a metallurgical sense. The nett result of polygonisation is the production of a recrystallised network of polygonal grains derived from the sub-grains of pre-existing grains. Thus it would appear that in these rocks pre-existing quartz, with triple junctions, has been reworked and incipient recrystallisation textures can be seen.

Plagioclase felspar occurs as granoblastic grains of an interlocking polygonal habit with straight boundaries, in places. However, it is largely interlobate, embayments of other felspar grains and quartz giving it irregular grain boundaries. Its average composition is sodic andesine, as determined by the Michel-Levy method, although grains with sodic labradorite properties are present in some specimens. The grains usually show gradational zoning, the cores being more calcic than the margins. In some cases, the plagioclase is antiperthitic and contains small aligned, rectangular bodies of untwinned alkali felspar which may show incipient microcline

twinning. This exsolution of alkali compent from a solid solution in plagioclase is generally accepted as indicating waning temperature placing the solid solution in disequilibrium. Sen (1959), however, has suggested that high pressures decrease the solubility of alkali felspar component in plagioclase, so that increasing pressure might equally account for the phenomenon in highly alkaline, high temperature plagioclases. Some of the plagioclase grains are untwinned and contain aligned rutile needles. However, progressive growth of spindle-shaped albite twins can be observed, from incipient, very fine ones until they are prominent. Finer discontinuous pericline twins show at right angles to the albite twins and produce a reticulate pattern with them, often similar to microcline twinning. In metamorphic rocks, twins may be of three origins; primary and of pre-metamorphic origin, primary and due to growth or recrystallisation, or secondary and due to post-crystallisation deformation. Primary growth twins are straight and often change width or terminate abrubtly within a crystal. Although deformation twins may show this property, they are usually thin and tapering, a result of reduction in lattice distortion near their ends, and are only sub-parallel. Also they usually show a much greater density of distribution than primary growth twins (Spry 1969). The fact that the twins described progressively develope in untwinned plagioclase is strong evidence that they are a product of deformation.

Clear, untwinned alkali felspar, with exsolved quartz blobs is present in quantities less than 1 per cent. It has developed microcline twins in sectors adjacent to angularities and tumescences on grain boundaries, probably as a result of stress. Small amoeboid pools of microcline enclose other grains, and appear at triple junctions amongst other felsic minerals. They contain numerous perthitic ribbons and exsolved quartz blobs.

Elongate aggregates of subhedral mafic grains define a rough fabric in thin section. The dominant minerals are pale pink to green pleochroic orthopyroxene, pale green clinopyroxene, greenish-brown pleochroic hornblende and dark red-brown biotite.

Orthopyroxene occurs as 1 mm polygonal grains showing triple junctions and straight boundaries indicative of equilibrium, and also as larger 2 mm embayed grains. They sometimes show undulose extinction and fine polysynthetic twins, possibly caused by strain. Replacement of orthopyroxene by pale yellow serpentine-like material is apparent along microfractures. This serpentine does not penetrate a mantle of blue-green amphibole and ore

granules often surrounding the orthopyroxene, and apparently replacing either orthopyroxene, or serpentine, or both. Himmelberg and Phinney (1967) describe a similar phenomenon developed in partly retrogressed granulite facies rocks from Wyoming. Some orthopyroxene is intimately associated with small biotite flakes which have developed from it via the blue-green amphibole fringe. Most grains of orthopyroxene seen are clouded with small ore granules and their outline is delineated by fine aggregates of ore. The larger embayed grains are skeletal, having pools of plagioclase and quartz, perhaps indicating their original formation as symplectitic intergrowths replacing a pre-existing mineral or forming by an early phase reaction.

Pale green, non-pleochroic clinopyroxene occurs in some rocks as equilibrated grains. It often partially encloses brown-green hornblende, and sometimes shows equilibrium relations to it. Clinopyroxene has been replaced by green secondary hornblende with which it is intergrown, quartz intervening as one of the reaction products.

Sparse brown-green hornblende, often full of ilmenite grains and dust, is seen to exist as euhedral crystals in equilibrium with ortho- and clinopyroxenes. It tends to be restricted to the darker bands in the gneiss, possibly as a result of their composition or possibly since dehydration did not proceed to completion in mafic bands at granulite facies, for some reason. It appears to have been transformed to green hornblende in which ilmenite has nucleated as discrete grains. It is partially replaced by biotite. Quartz-sieved secondary green hornblende is present as polygonal aggregates in some mafic bands, possibly formed by a combination of transformation of primary metamorphic hornblende and hydroxylation of pyroxenes.

Highly pleochroic biotite of a red-brown colour is sparse and exists as a replacive mineral, discrete flakes and as mantles to ore. The last two may indicate some recrystallisation of the first. It is partly replaced by chlorite.

Accessory minerals in these rocks include euhedral apatite, ilmenite, zircon and red rutile.

The early equilibrium assemblages represented by these quartz-and plagioclase-bearing gneisses are:

orthopyroxene-hornblende-alkali felspar
orthopyroxene-hornblende-clinopyroxene-alkali felspar
orthopyroxene-clinopyroxene-alkali felspar.

Later phases are serpentine, blue-green amphibole, green sieved hornblende, biotite and chlorite.

Migmatitic gneiss.

Strongly banded and well foliated migmatitic gneisses occupy the largest area on the islands. They contain inhomogeneously distributed hornblende and biotite which define a planar fabric in both light and dark layers parallel to the banding. Migmatitic gneisses occupy broad continuous belts up to 6 kilometres wide separated by prominent metasedimentary belts. Their banding varies from a few centimetres to decimetres thick, and is defined by pale felsic layers and dark layers rich in ferromagnesian minerals. The layering can exhibite very regular banding as at Traigh Halum on Coll (Plate 13a). The sharpness of layer boundaries seems to be more marked where layers are strongly contrasted compositionally. Banding is best developed in gneisses of average intermediate composition, where rapidly alternating leucocratic and melanocratic layers are equal in quantity. More acid gneisses show less regularity and banding is more gradational. There is a gradation in most cases from continuous mafic bands through discontinuous stripes to cases where the mafic material exists as boudinaged elongate lenses. In acid gneiss, biotite plays a larger role in the mafic bands than in intermediate gneiss, and is the predominant mafic mineral in the light bands, where it may be concentrated in selvedges a few millimetres thick, along which the gneiss will split. Mehnert (1968) calls such layered rocks stromatic migmatites.

A prominent and common feature of the acid migmatitic gneisses in the presence of discontinuous granular quartz-felspar veins, concordant with the banding. They may make up a small proportion of the gneiss, increasing until they become the dominant part of the rock (Plate 6d) in the most acid gneisses. In some cases the veins have discordant apophyses previously described in Chapter 2. At outcrop the veins stand out as ribs and because of encrusting lichens and the massive weathered surface of the gneiss, are frequently the only guide to foliation trends. Such veined migmatites have been termed phlebites by Mehnert (1968).

As is typical of migmatites (Mehnert 1968) those of Coll and Tiree contain numerous discordant quartzo-felspathic segregations and veins. They are similar to replacement pegmatites described by Ramberg (1956), and

often have a speckled appearance caused by large hornblende porphyroblasts. They are diffuse and often associated with boudinage, suggesting they crystallised preferentially in low pressure sites provided by cicatrices. They commonly occupy the middle limbs of a variety of asymmetrical folds, as described in Chapter 2. Whether they encouraged folding, being of low viscosity relative to the foliated gneiss, or were emplaced preferentially because of pressure considerations, is an open matter.

The thin melanocratic bands in the acid migmatites and the granular felsic veins often show pinch and swell structure and boudinage (Plate 5d), especially where the foliation is well developed as a result of deformation. The foliation in the gneiss is moulded around these irregularities giving rise to a phacoidal structure, biotite rich selvedges being discontinuous.

In strongly deformed gneisses, as in the intervening band in the Hogh Bay - Breachacha metasediment belt, an augen-like structure can be discerned, where the rock is made up of irregular phacoids of quartz and felspar around which thin mafic streaks are moulded (Plate 13b). This is unlike true augen or opthalmitic structure, as reviewed by Mehnert (1968), in which the "eyes" are single crystals. The "eyes" in the migmatites of Coll and Tiree are unlike similar structures described by Mehnert, in that they are not connected to discordant pegmatoid veins.

In general, the state of deformation and the relative roles played by banding and penetrative mineral fabric in the definition of the gneisses' foliation, is extremely varied, even over short distances along and across the strikes. Because of this it has proved impossible to distinguish between the rapidly and subtly varying types in lithological mapping. The dominant type is strongly quartzo-felspathic or acid.

Felsic minerals tend to be polygonally granoblastic in mafic bands and interlobate to amoeboid granoblastic in leucocratic bands.

Quartz is in grains up to 1 cm in diameter, of an amoeboid nature with sutured boundaries in some cases. It encloses other minerals and occurs as inclusions and exsolved blobs in others. Most grains exhibit strong undulatory extinction, the subgrains having fairly straight boundaries and each subgrain having uniform extinction. This would seem to indicate that polygonisation has progressed to the state where dislocations have migrated to form planar arrays and a state of equilibrium is being approached in the quartzes. In some felsic bands 1mm polygonal aggregates of quartz have

recrystallised indicating attainment of equilibrium. They show varying degrees of strain, from strain-free to slightly undulose.

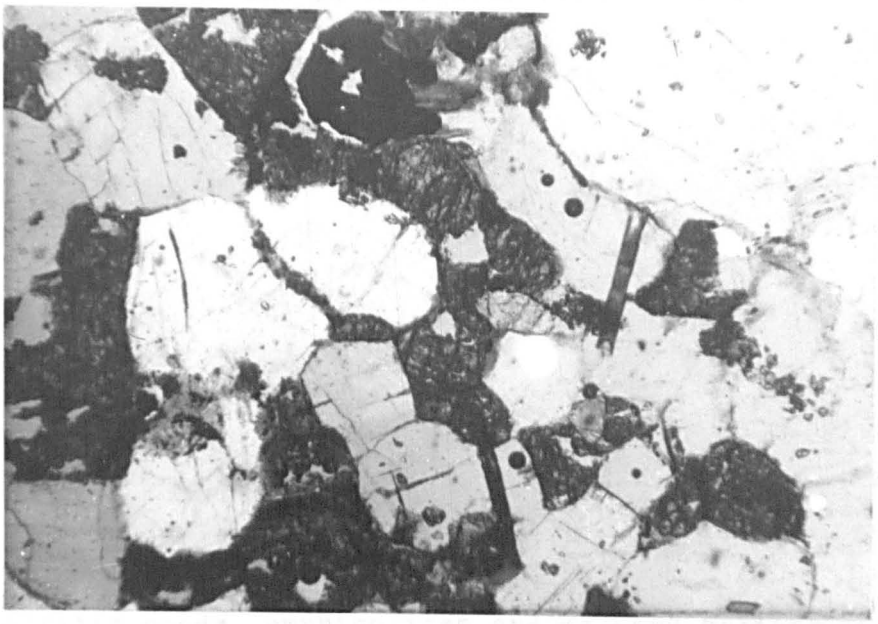
Interlobate plagioclase grains up to 1 cm across are often fresh and show prominent albite-type twins which are somewhat spindle-shaped but tend to be better developed than in the hypersthene-gneisses. Checkerboard twins are present, made up of parallel albite twins and finer discontinuous pericline twins at right angles. The composition of the plagioclase varies from An_{24} to An_{38}, averaging about An_{28}. Grains are usually clear but may be clouded and saussuritised, the alteration being late stage.

Untwinned alkali felspar contributes a variable proportion to the mineral assemblages, being clear to sericitised. It contains exsolved blobs of quartz. Irregularly distributed microcline twinning is sometimes seen, being concentrated near angularities in grain boundaries. The twins are probably indicative of strain disrupting the alkali felspar lattice. Where microcline is well developed, myrmekite, consisting of blebs and small tubes of quartz in a sodic plagioclase host, is seen as irregular interlobate grains. The formation of myrmekite in gneisses has been explained as due to exsolution of myrmekite component ($CaAl_2Si_6O_{16} + NaAlSi_3O_8$) from alkali felspar formed at high temperatures, at alkali felspar grain boudaries, in response to waning temperature (Carman and Tuttle, 1963; Hubbard, 1966). Textural evidence for other modes of formation of myrmekite is not apparent.

The most common mafic mineral in the migmatitic gneisses is pleochroic pale yellow to brown biotite as individual flakes up to 2 mm in length and as felted masses of similar grain size, defining a strong foliation. It is also found as a fine grained symplectitic intergrowth with quartz where it replaces green hornblende (Plate 12b), and as small flakes enclosed by hornblende together with small quartz grains. In addition it is developed as a partial radiating fringe to ore grains. Commonly, small laths of prehnite occur along the cleavage of biotite.

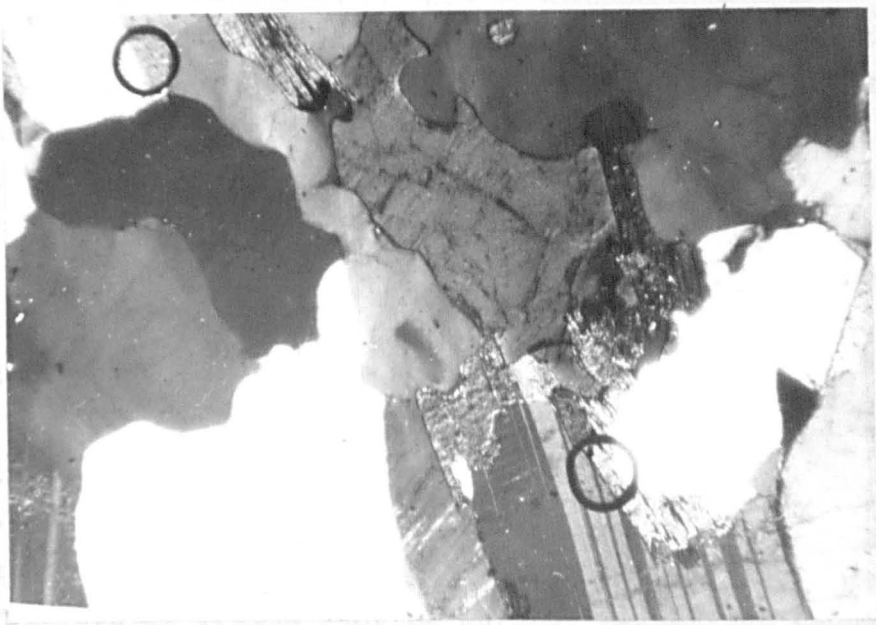
Green hornblende is most common in the melanocratic bands where it is dominant over biotite, and exists as polygonal aggregates exhibiting triple junctions. It frequently includes blobs of quartz. In the acid layers it is usually present as discrete crystals often partly replaced by biotite/quartz intergrowths and small granules of green hornblende are scattered in masses of the intergrowth.

Accessory minerals found in the migmatitic gneisses include apatite, zircon, ore, sphene, epidote, prehnite and chlorite. Sphene occurs as



a

Photomicrograph of hypersthene-gneiss, showing relic polygonal texture and orthopyroxenes partly altered to biotite and mantled by blue-green amphibole.
(Spec. 78SW1A, p.p.l., x 60)



b

Photomicrograph of nebulitic gneiss, showing interlobate texture, and hornblende replaced by a biotite/quartz intergrowth.
(Spec. 4931, x.p., x 60)

brown lozenge-shaped grains and as irregular rims to ore. Its presence is analogous to that of rutile in the hypersthene-gneiss, rutile being the titanium mineral typical of the granulite-facies and sphene of the amphibolite facies.

The mineral assemblage represented by the migmatites is:

quartz-plagioclase-hornblende-biotite-(alkali felspar).

Nebulitic gneiss.

Within the dominant well foliated migmatitic gneisses are to be found wide belts up to 500 metres across and narrow diffuse bands of unbanded acid gneisses of a pale pink colour. They are massive and almost devoid of foliation, although containing hornblende and biotite. Relics of pre-existing melanocratic bands, amphibolites and metasedimentary rocks are seen as rotated rafts, enclaves, schlieren and pods, often containing a foliation describing fragments of folds (Plates 6a, 13c). The schlieren provide the only guide to foliation trends within outcrops of this rock type. Noteable belts of nebulitic gneiss occur near Balephetrish Hill and along the eastern coast of Tiree (Map 3), from Rudha Hogh to Fasagh, Bernera and from Loch Ronard to Friesland on Coll (Map 1).

The rafts and enclaves represent dislocated, boudinaged competent bands, which have been further deformed to elongate, partly rodded schlieren. The schlieren themselves are frequently boudinaged, the process resulting in numerous black pods of hornblendite (Plate 13d). Some metasomatic exchange would seem to have taken place during deformation between the fragments and the nebulitic gneiss, as the former become progressively darker with their increasingly deformed state.

The nebulitic gneisses are coarser and more granoblastic than the banded hornblende-biotite-migmatitic gneisses, although having a similar mineral content. The predominant felsic minerals are seen as a complex interlocking network of irregular grains showing curved, re-entrant boundaries and only rare triple junctions. The ferromagnesian minerals are randomly oriented for the most part, although some sections display a weak fabric.

Quartz is prominant as amoeboid grains up to 8 mm in diameter. It has undulose extinction, indicative of polygonisation perhaps in a less constrained environment than that of the migmatites. Smaller grains of quartz

PLATE 13

Plate 13

a. Banded migmatitic gneiss: Traigh Halum, Coll.

b. Quartzo-felspathic augen in strongly deformed gneiss: Clabhach, Coll.

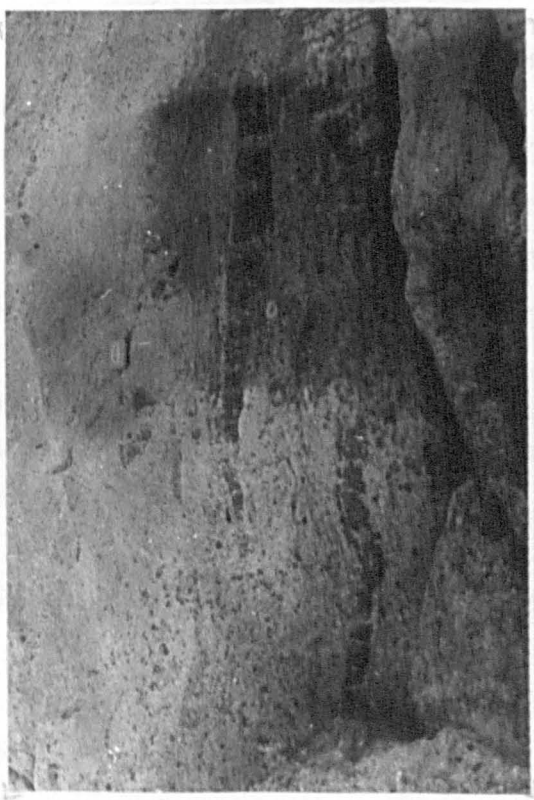
c. Enclaves of calc-silicate rock basic material in nebulitic gneiss: Ballyhogh, Coll.

d. Boudinaged mafic schlieren in nebulitic gneiss: Friesland, Coll.

a

b

c



d

are associated with microcline pools, as granular aggregates showing little evidence of post-crystalline strain, straight grain boundaries and sometimes triple junctions. This would seem to indicate that polygonisation of quartz has proceeded to completion, subgrains of the large strained grains having recrystallised in association with microcline.

Plagioclase, ranging in composition from An_{25} to An_{35} is in regular grains, interlobate to some extent but showing a few straight boundaries and triple junctions. It is unzoned and has well developed albite twins, only slightly spindle shaped. Frequently, the plagioclase is antiperthitic, containing rounded cuneiform microcline pools, aligned parallel to cleavage and twinning. Plagioclase has often been clouded and saussuritised at a late stage.

Sparse interlobate grains of largely untwinned alkali felspar are found, with similar properties to those in migmatitic gneisses. Amoeboid pools of microcline fill interstices in the framework of the other minerals, and range in size from 0.5 mm to 1 cm in diameter. The larger pools are perthitic, exsolved sodic felspar being in aligned ribbons and patches. As in the migmatites, where microcline abuts plagioclase, myrmekite is frequently seen.

Biotite in large flakes, up to 5 mm in length, and felted aggregates is randomly distributed. It is pleochroic from yellow to pale brown. Some biotites have interleaved prehnite wedges, others are greenish, being altered to chlorite and dusty iron ore. The occurrence of prehnite wedges in biotite is possibly a manifestation of later greenschist facies metamorphism, prehnite developing preferentially along the sheet lattice of biotite and taking up calcium released by the breakdown of plagioclase.

Sparse, green, subhedral, unsieved hornblende, which is unaligned and frequently partially replaced by a biotite/quartz symplectite (Plate 12b).

Accessory minerals are apatite, zircon, iron ore, sphene, epidote, prehnite and chlorite.

The mineral paragenesis of the nebulitic gneisses is identical to that of the banded migmatitic gneisses.

Diopside-gneiss.

Diopside-gneisses are associated with nebulitic gneisses where they enclose metasedimentary bands, or are adjacent to belts of metasediments,

and sometimes at the junction between metasediments and banded migmatites. They are phacoidally foliated, showing augen of quartz and felspar, and porphyroblasts of pale green diopside and darker green amphiboles. The 5 km long boundary between the gneisses of the Rudha Hogh-Fasagh area, which are largely nebulites, and the Hogh Bay - Breachacha metasedimentary belt is marked by a continuous layer of diopside gneiss some 10 m wide. In it the proportion of diopside culminates in a metre thick band close to the nebulite proper. Similar rocks form a thin mantle to the strongly deformed hornblende-biotite-gneiss band within this metasediment belt, and at its western boundary where they are interleaved with thick granofels bands. A similar relationship exists at the east side of Ben Feall. Diopside-gneiss forms a discontinuous 20 m thick band within the nebulitic gneisses at Fasagh.

The field relations of these enigmatic rocks suggests that they are connected with formation of nebulitic gneiss adjacent to metasedimentary bands, and they could have formed by metasomatic activity associated with the partial mobilisation of the nebulites.

A complex textural pattern is present in diopside-gneisses. Grains of felsic minerals vary in size and outline. Mafic minerals are often poikiloblastic, though biotite defines a penetrative foliation. Quartz is in amoeboid grains up to 1 cm across, sometimes elongate parallel to the fabric. Subparallel strain bands are typical in these. Polygonal aggregates of recrystallised quartz associate with microcline grains.

Partly saussuritised and clouded plagioclase (An_{30} to An_{45}) has a polygonal habit with triple junctions (Plate 14a) and has been recrystallised. It shows regular albite twins without deformational features. Myrmekite is rare although its usual companion in gneisses, microcline, is the dominant felsic phase.

Alkali felspar is represented by perthitic microcline with strong "tartan" twinning. Grains are up to 3 mm in extent often amoeboid, and usually in association with and embaying irregular quartz grains. Throughout much of the rock aggregates of polygonally granular microcline (0.5 mm grain size) and quartz form a matrix to coarser ferromagnesian minerals. Textures in these aggregates are consistent with recrystallisation. Microcline appears to increase in importance with diopside.

Pale green diopside, devoid of exsolution lamellae, is in skeletal

grains up to 5 mm across. It never shows a euhedral form, and encloses pools and irregular blobs of quartz, microcline and plagioclase (Plate 14a). Its poikiloblastic form is a disequilibrium texture, and as it is unevenly distributed at outcrop, it may have grown under metasomatic conditions after the development of fabric in the gneiss. As its distribution is followed by microcline, some relationship may have existed between the minerals during their growth.

Pale to blue-green amphibole occurs as skeletal, coarsely quartz-sieved, poikiloblasts. It is also intergrown with diopside, quartz intervening between the two phases. Each discrete area of amphibole within a diopside grain is in optical continuity with the others, and in certain sections with diopside itself. Such a situation is usually caused by epitaxic intergrowth and is interpreted as due to replacive nucleation and growth of amphibole being controlled by the pre-existing diopside lattice. Pashley (1956) has shown that substrate and overgrowth may differ considerably in structure and bonding, and still show epitaxic relationships.

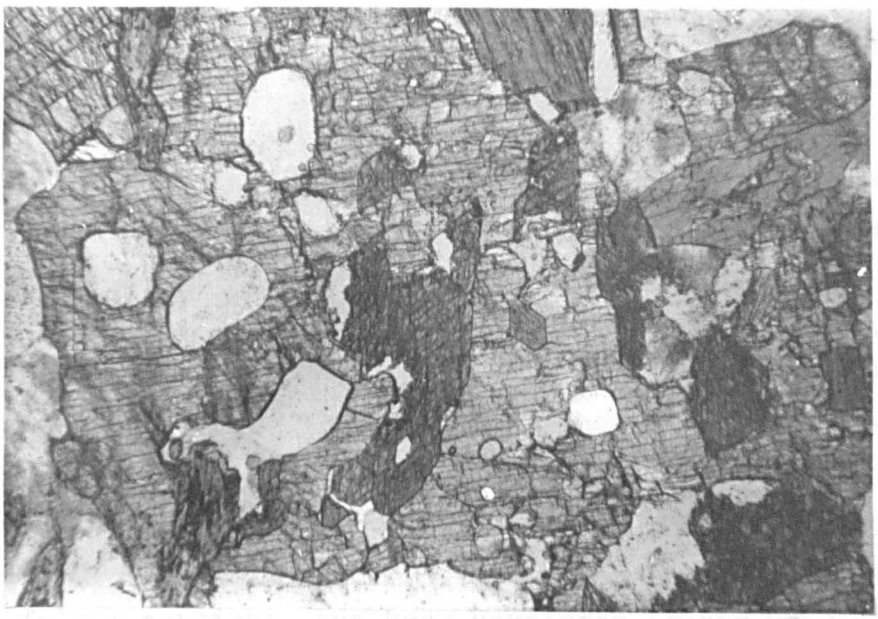
Prehnitised and chloritised straw yellow biotites define a planar fabric.

Accessory minerals are apatite, zircon, ore, epidote, chlorite and prehnite.

Epidote-chlorite-gneiss.

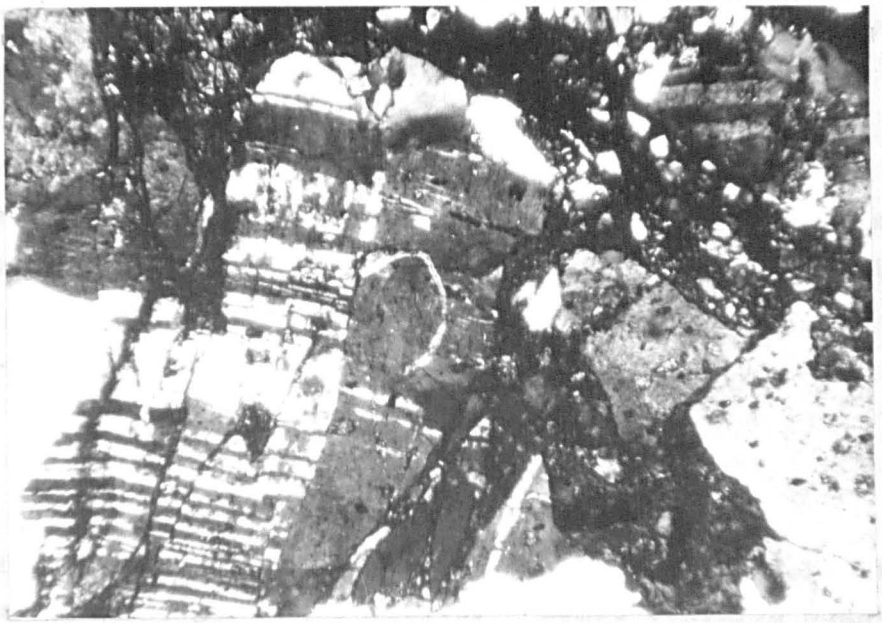
Pink, green and red sheared gneisses are found in localised belts, generally along F_6 shear belts. Their foliation is stronger than those of other gneisses and is defined by chlorite flakes. The concordant quartz-felspar veins are foliated too. Pre-existing foliation is disrupted, wedges and lenses of banded gneiss being surrounded by finely foliated material. In the shear belts there is a transition from grey hornblende-biotite-gneiss through a pink disrupted and chloritic gneiss to a final product of finely foliated gneiss with a greenish tinge. The new foliation (S_6) is phacoidal and full of streaked out felsic lenses. The most prominent development of epidote-chlorite-gneiss is to be found at the NE part of Coll, and following an F_6 shearbelt from Traigh Bousd to Gallanach (Maps 1 and 2) along the NW shore.

All phases examined in thin section show signs of strong deformation. Quartz, in interlobate plated grains, shows extreme strain shadowing



a

Diopside poikiloblast in diopside-gneiss, showing included quartz and felspar, and epitaxically intergrown and replacive hornblende.
(Spec. 7321I, p.p.l., x 60)



b

Fractured bent and granulated quartz and felspar in epidote-chlorite-gneiss. Patches of chlorite pseudomorphing biotite and hornblende are shown together with clusters of high relief epidote.
(Spec. 56916E, x.p., x 60)

in sub-parallel bands emanating from tumescent boundaries, and inclusions. Grain boundaries show granulation on a scale of 0.01 mm. Between some of the larger quartz grains can be found fine polygonal recrystallised aggregates of quartz and feldpar.

Plagioclase is strongly clouded, and twinning therefore indistinct. Twins are seen to be curved, kinked and displaced by microfractures (Plate 14b). The alteration products are mainlyvery fine grained, but sericite, epidote, clinozoisite and pools of clear albite can be discerned.

Microcline is prominent showing bent and fractured polysynthetic twins. It is largely unaltered but sometimes encloses irregular flakes of muscovite. Chlorite and ore dust pseudomorph biotite in most cases, and the original fabric is preserved. However in the most strongly deformed gneisses, a new penetrative planar fabric, controlled by thin films of chlorite separating granular felsic portions, has been developed (Plate 14b). Chlorite has also replaced hornblendes, first showing as thin streaks in the amphiboles, parallel to the new foliation. It has culminated complete replacement and transposition of earlier hornblende into foliation aligned streaks. Altered hornblende frequently has a sphene rim.

Epidote grains occur in clusters, not apparently having replaced any earlier phase but nucleated in favourable positions. These are common in altered plagioclase grains.

Prehnite occurs in discordant calcite-quartz veins. Other accessories include apatite ore and sphene.

The assemblage represented by growth phases in the epidote-chlorite-gneiss, discounting the disappearing phases plagioclase, microcline, biotite and hornblende, is:

quartz-albite-muscovite-epidote-chlorite.

Discussion.

The petrographic evidence from the hypersthene-gneisses shows that they crystallised in the granulite facies. Relic textures suggest that equilibrium existed betweeen the phases present. Brown hornblende in equilibrium with the two pyroxenes could have been the result of incomplete, and varying, dehydration of the rocks by the close of the early granulite facies event.

The succeeding metamorphic history of the quartzo-felspathic gneisses

is largely confined to the amphibolite facies, and shows the effects of the increasing geochemical activity of hydrous fluids.

The transition from granulite facies to amphibolite facies is shown by the growth of green hornblende from earlier brown hornblende and clinopyroxene. Although direct evidence is not seen here, Himmelberg and Phinney (1967) suggested that green hornblende forms at the expense of orthopyroxene via blue-green cummingtonite mantles following the reaction:

$$\underset{\text{cummingtonite}}{3(Mg,Fe)_7Si_8O_{22}(OH)_2} + \underset{\text{anorthite}}{14CaAl_2Si_2O_8} + 4H_2O \rightleftharpoons$$

$$\underset{\text{green hornblende}}{7Ca_2(Mg,Fe)_3Al_4Si_6O_{22}(OH)_2} + \underset{\text{quartz}}{10SiO_2}$$

during retrogression of granulite facies gneisses.

Initially biotite developed by replacement of brown hornblende and orthopyroxene. Further biotite crystallisation, with green hornblende, followed in the amphibolite facies migmatitic gneisses. Yet a further development of biotite occurred at the expense of green hornblende when it was replaced by biotite-quartz symplectites. The potassium essential for the latter transformation may have been of external origin or resulting from the exsolution of alkali felspar component from plagioclase. There is no evidence in amphibolite facies gneisses for the breakdown of alkali felspar.

The marked disequilibrium textures in the felsic minerals of some migmatitic gneisses is succeeded in the nebulitic gneisses by signs of small scale recrystallisation. Microcline is prominent in the amphibolite facies gneisses, and plagioclase-microcline interfaces provided sites for the development of myrmekite. Whether microcline growth was in response to changes in alkali felspar structure or exsolution from high temperature plagioclases during the amphibolite facies or due to influx of alkaline fluids can only be established by geochemical means.

In the epidote-chlorite-gneisses, biotite and hornblende have been extensively replaced by chlorite and prehnite, and epidote has grown in altered plagioclase grains leaving relics of a sodic nature. Calcite-prehnite veins are common and muscovite has partly replaced microcline. The presence of the assemblage quartz-albite-epidote-chlorite-muscovite with prehnite shows the effects of a late stage greenschist facies metamorphism (Turner and Verhoogen, 1960). The widespread presence of prehnite in all the gneisses is probably a manifestation of this event.

At this point it is convenient to discuss the evolution of some of the

is largely confined to the amphibolite facies, and shows the effects of the increasing geochemical activity of hydrous fluids.

The transition from granulite facies to amphibolite facies is shown by the growth of green hornblende from earlier brown hornblende and clinopyroxene. Although direct evidence is not seen here, Himmelberg and Phinney (1967) suggested that green hornblende forms at the expense of orthopyroxene via blue-green cummingtonite mantles following the reaction:

$$\underset{\text{cummingtonite}}{3(Mg,Fe)_7Si_8O_{22}(OH)_2} + \underset{\text{anorthite}}{14CaAl_2Si_2O_8} + 4H_2O \rightleftharpoons$$

$$\underset{\text{green hornblende}}{7Ca_2(Mg,Fe)_3Al_4Si_6O_{22}(OH)_2} + \underset{\text{quartz}}{10SiO_2}$$

during retrogression of granulite facies gneisses.

Initially biotite developed by replacement of brown hornblende and orthopyroxene. Further biotite crystallisation, with green hornblende, followed in the amphibolite facies migmatitic gneisses. Yet a further development of biotite occurred at the expense of green hornblende when it was replaced by biotite-quartz symplectites. The potassium essential for the latter transformation may have been of external origin or resulting from the exsolution of alkali felspar component from plagioclase. There is no evidence in amphibolite facies gneisses for the breakdown of alkali felspar.

The marked disequilibrium textures in the felsic minerals of some migmatitic gneisses is succeeded in the nebulitic gneisses by signs of small scale recrystallisation. Microcline is prominent in the amphibolite facies gneisses, and plagioclase-microcline interfaces provided sites for the development of myrmekite. Whether microcline growth was in response to changes in alkali felspar structure or exsolution from high temperature plagioclases during the amphibolite facies or due to influx of alkaline fluids can only be established by geochemical means.

In the epidote-chlorite-gneisses, biotite and hornblende have been extensively replaced by chlorite and prehnite, and epidote has grown in altered plagioclase grains leaving relics of a sodic nature. Calcite-prehnite veins are common and muscovite has partly replaced microcline. The presence of the assemblage quartz-albite-epidote-chlorite-muscovite with prehnite shows the effects of a late stage greenschist facies metamorphism (Turner and Verhoogen, 1960). The widespread presence of prehnite in all the gneisses is probably a manifestation of this event.

At this point it is convenient to discuss the evolution of some of the

macroscopic features in the quartz-felspathic gneisses, in the light of petrographic evidence.

In the granulite facies hypersthene-gneisses the presence of foliation, lineations, banding and diffuse pegmatoid bodies is plain evidence for a metamorphic history before the high grade event, perhaps equally as complex as the succeeding phases of retrogression. Mineralogical banding could have developed initially at any stage during this early history and may have been modified further. However, it is generally accepted that metamorphic banding is a product of amphibolite facies conditions when chemical mobility and tectonic activity would be strong enough to effect the element redistribution necessary (see Winkler, 1967). The relatively homogeneous nature of the Ben Hynish hypersthene-gneisses perhaps indicates strong recrystallisation at some later stage, probably at granulite facies. Their homogeneity reflects their more acid composition compared with well banded intermediate hypersthene-gneisses from elsewhere on Tiree. The presence of discordant pegmatoid bodies with high grade assemblages in the hypersthene-gneisses shows the activity of silicate fluids after the formation of the banding. As they are cut by basic dykes with granulite facies mineralogy, they were formed before granulite facies and owe their present mineralogy to high grade metamorphic processes rather than primary crystallisation.

Before discussing the development of the gneisses during retrogressive metamorphism it must be said that it is assumed that the present lower grade quartzo-felspathic gneisses of Coll and Tiree are the downgraded products of an earlier series of granulite facies gneisses. The following facts justify this assumption to some extent:

i) No structural or petrographic evidence is seen for two gneiss complexes distinct in their origins.

ii) Well banded orthopyroxene-bearing gneisses grade into migmatitic gneisses on Tiree.

iii) Both granulite facies and amphibolite facies gneisses are cut by metabasite intrusions which will be shown to have a common origin prior to granulite facies metamorphism.

Some of the features of the migmatitic gneisses, such as the strong interbanding of the felsic and mafic portions in intermediate gneiss and the presence of concordant granitoid veins, may well have been inherited from granulite facies gneisses, having lost their high grade assemblages by retrogression. On the other hand they may have formed by wholesale

reworking of the earlier banding during D_1. The solution of the dilemma hinges on the age of the formation of F_1 intrafolials, which is a problem in itself (p. 22).

The major modifications to the migmatitic gneisses seem to have been caused largely during D_3. Regular mafic bands give way to discontinuous stripes and lenses as a result of boudinage. Digestion of mafic layers in some types of migmatite may have resulted in the crystallisation of biotite-rich selvedges under conditions of deformation. Granitoid veins in acid migmatites have been boudinaged, and in places augen gneisses have been formed by this process having proceeded to extremes.

Homogeneous nebulitic gneisses seem to have formed during D_3, in areas of very acid gneiss. Their homogeneous nature indicates the extent to which their fabric was lost by chemical mobility, whilst their content of flattened mafic fragments, boudinaged schlieren, and hornblendite pods shows that deformation was an equally important factor contributing to their formation. The ionic mobility in the nebulitic gneisses is emphasised by the progressive basification of mafic fragments within such belts, and the marginal formation of metasomatic diopside-gneisses in favoured localities. Nebulitic gneisses, considered in bulk, are very acid, basic material being only in small quantities. This factor may well have a bearing on their formation. An acid gneiss, with only small proportions of mafic bands, would tend to act as a focus for migration of hydrous fluids. More prominent mafic bands elsewhere, in the migmatitic gneisses, would act as barriers to diffusion and so protect the associated felsic bands to some extent. The availability of migrating hydrous fluids would actively contribute to mobilisation of the gneiss, perhaps by effecting supercritical solution of some portions, or even partial melting if temperature was high enough. In any case, the viscosity of the nebulitic gneisses was lowered relative to adjacent rocks, bolstered by their larger proportions of interlayered mafic bands, and so they deformed most easily during D_3. The chemical activity of the migrating fluids must have enabled the gneiss to become homogenised, and contrasted lithologies within and adjacent to it were able to take part in a metasomatic exchange process with the nebulitic gneiss.

The progress of events, culminating in the formation of nebulitic gneiss, is in many respects similar to that described by Myers (1970) from Harris, where the ultimate product of "late Scourian" gneissification was a homogeneous granitic gneiss. Myers has shown that the major gneiss form-

ing process in the Lewisian of Harris left a mixed series of genetically linked gneiss types, which were subsequently altered by alkali metasomatism during the Laxfordian to the present Harris migmatite complex. The Laxfordian left little tectonic imprint on the earlier gneisses.

Mobile fluids, either silicate melts or supercritical, hydrous silicate solutions, coalesced in both migmatitic and nebulitic gneisses to form discordant pegmatoid bodies, some of which were intimately associated with F_4 and F_5 minor folds.

The late greenschist facies modifications to the gneisses seem to have been largely accomplished by tectonic means during D_6. Some water must have entered to give chloritic assemblages. // Some points of interest arising from this petrographic study of quartzo-felspathic gneisses can be summarised thus:

a) Three distinct metamorphic phases at granulite, amphibolite and greenschist facies can be recognised.

b) There were already gneissose rocks before the granulite facies.

c) During the amphibolite facies period a process of selective gneissification reworked the older granulite facies gneisses, culminating in the formation of nebulitic gneiss belts and associated metasomatic diopside-gneiss. The gneisses were almost completely changed to hydrous assemblages.

d) Three tectonic episodes largely effected the gneissification process:

D_1 - S_1 banding formed in the migmatitic gneisses

D_3 - S_1 deformed and nebulitic gneiss developed

D_6 - Gneisses sheared in steep belts.

Metasediments.

Rocks of metasedimentary aspect are widespread on Coll and Tiree in three distinctive types of location. Wide, continuous belts of variegated metasediments up to a kilometre in width are found on Tiree between Bhasapol and Ceann a' Mhara, Balephetrish Bay to Hynish and Vaul to Gott Bay. On Coll, similar belts outcrop from Feall Bay to Crossapol Bay, being repeated between Breachacha and Hogh Bay by the major isoclinal fold at Port na Luing, and from Clabhach to Gorton. Discontinuous bands up to 30 m wide can be mapped within the dominant quartzo-felspathic gneisses, as at Port na Luing, Arinthluic, SE of Loch an Duin and near Cranaig on Coll. Enclaves, lenses and pods up to 20 m across, but generally much smaller, are

widely distributed throughout the gneisses, being especially distinctive in nebulitic gneisses (Plate 13c).

Garnet-biotite-gneiss.

Garnet- and biotite-bearing quartzo-felspathic gneisses occur on all three scales on Coll and Tiree. They may be seen in a 500 m thick belt at Hynish, and at Scarinish as a 20 m thick band in the Vaul-Gott Bay belt on Tiree, in 20 m thick bands west of Arinthluic and SE of Loch an Duin, and as irregular refolded pods between Gallanach and Toraston on Coll (Fig. 9). They are easily recognisable in the field by their rusty weathered surface. They infrequently exhibit mineralogical banding caused by inhomogeneities in the distribution of garnet. Garnets occur as large porphyroblasts up to 5 cm across, around which the foliation is deflected. Felsic minerals often exhibit a blue opalescent lustre. The foliation is stronger than in the migmatitic gneisses, and often contorted. Prominent, massive garnetiferous quartzo-felspathic veins occur, concordant with the foliation, but free of fabric themselves. They are rarely more than a few decimetres in thickness, and are discontinuous along the strike. They bear a strong resemblance to garnetiferous leucosomes in the granofelses and some metabasites.

In thin section these rocks show a very varied mineralogy but all are foliated and contain garnet and biotite.

Quartz is usually co-dominant with plagioclase. Some specimens, notably from the band near Arinthluic on Coll, show well developed flat plates or "plattung" of quartz up to 2 cm long (Plate 15a), defining a strong foliation. The plates show polygonised lamellae perpendicular to their length. Between the plates, or throughout the rock in plattung-free specimens, large amoeboid grains are seen to be polygonised in an irregular fashion, and quartz embays, and is embayed by, plagioclase. Polygonal aggregates of quartz, up to 2mm in grain size, showing equilibrium relationships have recrystallised in some specimens, but their equilibrium has been disturbed, giving some curved grain boundaries. In heterogeneously deformed garnet-biotite-gneisses, quartz with felspar is found in phacoidal aggregates showing rotated triple junctions, and as fine, 0.5 mm, polygonal grains in the surrounding strongly deformed biotite-rich portions.

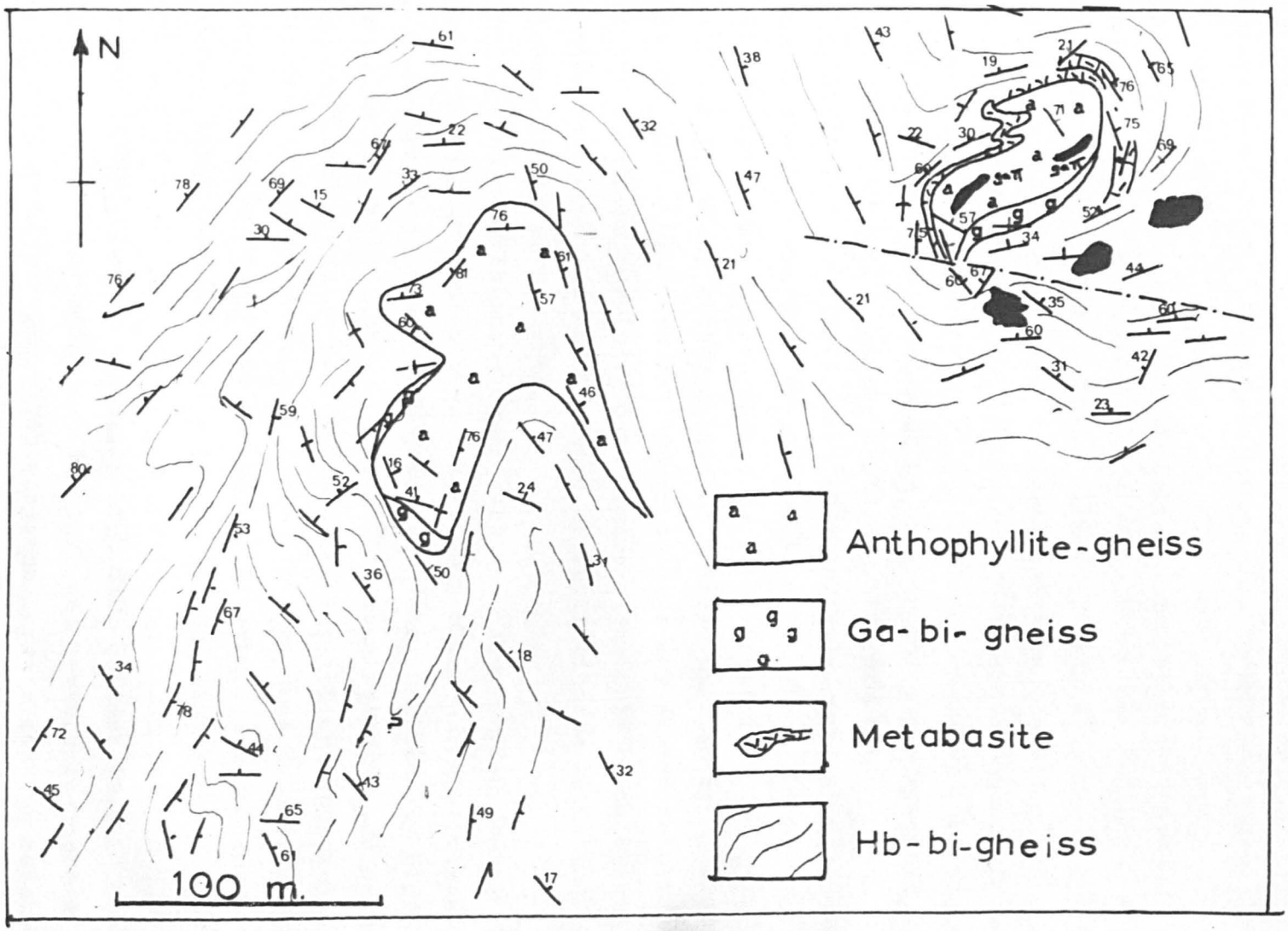


Fig.9: map of possible metasedimentary bodies
E of road between Gallanach and Torastan, Coll.

Plagioclase, from An_{22} to An_{52}, is the only felspar present in these rocks. They are untwinned only in orthopyroxene-bearing varieties, and generally show deformation twins to some degree. The untwinned grains and other twinned grains in orthopyroxene-bearing varieties, are polygonal and inequilibrium. This state of equilibrium, with respect to plagioclase, is disturbed in other types, the mineral showing an interlobate granoblastic habit, zoning with calcic cores, and associated myrmekite.

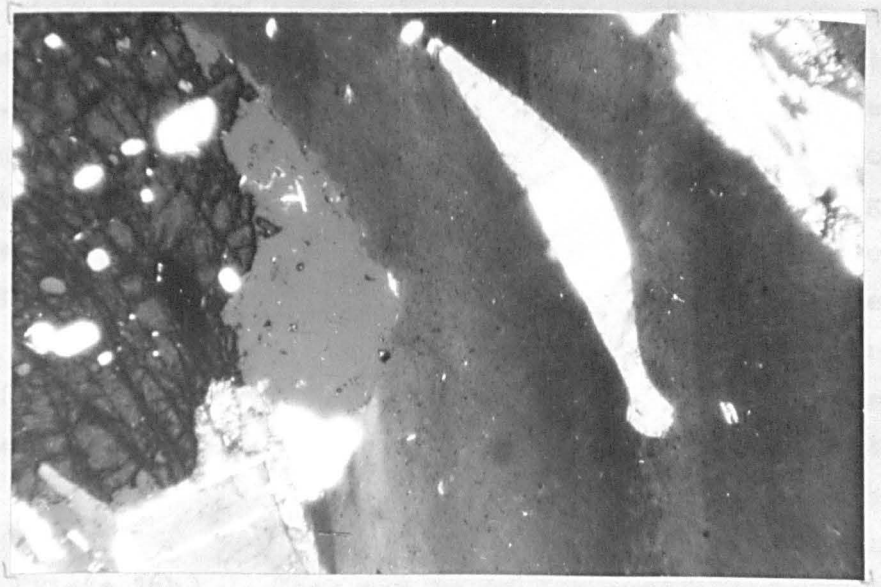
Al_2SiO_5 polymorphs are rarely found in garnet-biotite-gneisses, the most common being sparse laths of kyanite, within the foliation and not clearly related to other minerals. However, in a specimen from Rudha Craiginnis on Tiree, both kyanite and sillimanite are seen. Sillimanite occurs as acicular grains enclosed by deformed garnets which are streaked along the foliation (Plates 15b, 16a). The sillimanite needles define rotated trails indicating the existence of an earlier foliation overgrown by the garnet. Numerous small, stubby kyanite grains are aligned with the strong newer foliation. The significance of this isolated relationship will be covered in the discussion to this section.

Highly pleochroic orthopyroxene is to be found in a few specimens. Its habit is generally stubby subhedral prisms, mantled by blue-green amphibole and ore dust. It does not appear to have any direct relationship with garnet and reactions involving both minerals can only be inferred. It is often seen to be mantled and replaced by small biotite laths and green hornblende, sieved with quartz.

Pale green cummingtonite is present in a few samples, being replaced by a biotite/quartz symplectite. Similarly where garnet-biotite-gneiss abuts anthophyllite-gneiss, anthophyllite grains replaced by biotite/quartz are found in the garnetiferous rock. Both cummingtonite and anthophyllite could be amphibolite facies replacements for earlier orthopyroxene.

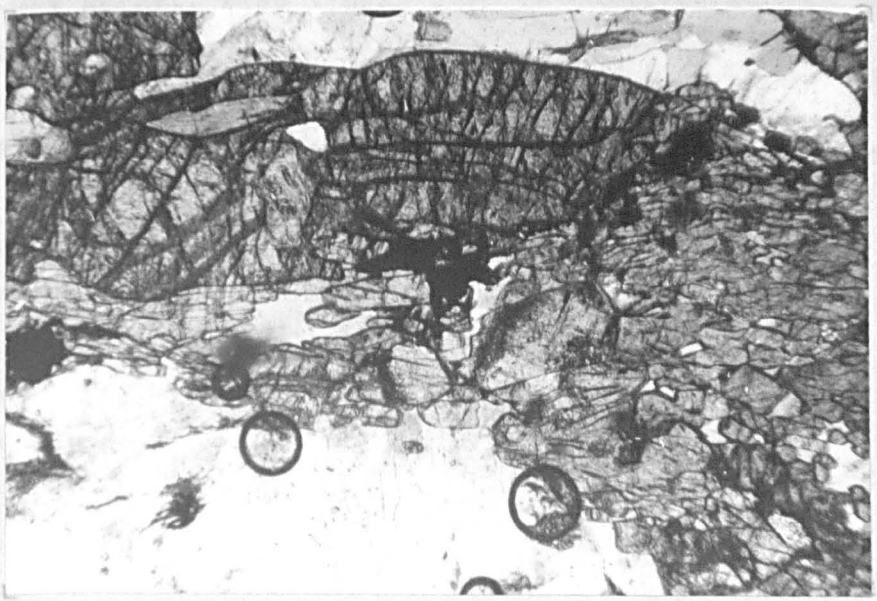
Biotite is always present in these rocks to some extent, usually being a strongly coloured "foxy" red variety. Texturally, its formation is multiphase, large flakes defining the foliation, smaller ones showing a multiplicity of replacive roles. A partial analysis of such a biotite from specimen 4965A has an unusually high TiO_2 content (Table 2).

Pink garnet is ubiquitous. Its form and texture suggests at least three generations of growth. The earliest garnets contain rotated helycitic trails of dust, sillimanite and biotite, and are often deformed along the present foliation (Plate 15b). They would seem to be syntectonic in



a

Plattung-type quartz with diagonal strain shadows, also sieved irregular garnet.
(Spec. 4965A, x.p., x 60)



b

Deformed early garnet aligned with foliation defined by kyanite.
(Spec. 64NW1B, p.p.l., x 60)

origin (Spry 1969). The most common garnets are up to 5 cm in diameter and have a skeletal to sieved habit (Plate 15a). They contain numerous inclusions, dominantly quartz blobs, but including biotite, plagioclase and amphiboles, usually all randomly oriented. There are rare examples of vague rotated trails in garnets of this generation. Their content of various inclusions and helicitic trails would suggest that they grew rapidly from a pre-existing ferromagnesian mineral or minerals under orogenic conditions. They often show an inclusion-free outer zone, either indicative of clearing of the inclusions or further sluggish growth involving diffusion of components from the surrounding matrix under static conditions. The outer zone usually has some planar faces developed in contact with and cutting across biotite flakes in the fabric. Smaller, euhedral, unsieved garnets up to 2 mm in diameter are found evenly scattered through the matrix of other minerals (Plate 16b). They may be of the same generation as the inclusion-free zones, and probably represent growth from nuclei during favourable conditions, a gradation of sizes up to 2 mm being present. Where biotite abuts garnet, the foliation is truncated and deflected around the garnets. Crystal faces of the garnets are rarely embayed, and the mineral appears to be in equilibrium with biotite, although in a few cases biotite appears to replace garnet, fine flakes with interleaved alkali felspar mantling the garnet.

Accessory minerals include ore, apatite, zircon, rutile and tourmaline.

Assemblages represented by these quartz- and plagioclase-rich rocks are:

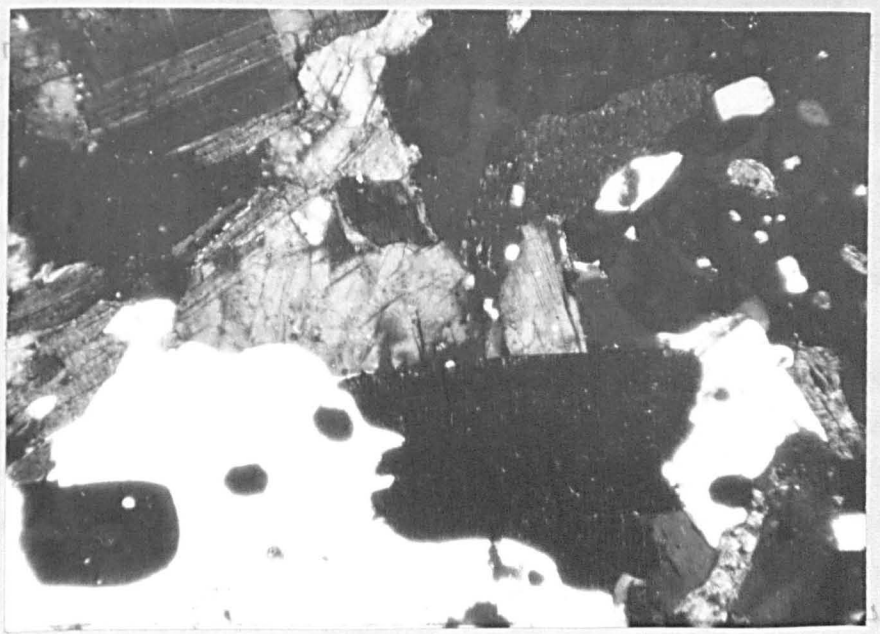
Orthopyroxene-garnet-biotite
garnet-biotite
garnet-biotite-kyanite
garnet-biotite-kyanite (~~-~~sillimanite)
garnet-biotite-cummingtonite
garnet-biotite-anthophyllite
garnet-biotite-hornblende.

Anthophyllite-gneisses.

Buff felspathic rocks, containing laths of a black amphibole which defines a strong linear fabric, are found in complexly folded pods up to 20 m across, SE of the road between Gallanach and Toraston (Fig. 9).

a

Trails of sillimanite needles within deformed garnet.
(Spec. 64SW5A, x.p., x 60)



b

Sieved large garnets, unveined small polygonal types, in garnet-biotite-gneiss.
(Spec. PLGR, x.p., x 60)

They are associated with garnet-biotite-gneisses, and are devoid of banding, being contorted on a small scale.

Petrographic examination emphasises the strong linear fabric of the anthophyllite-gneisses.

Quartz is absent as a primary mineral, only being found as exsolved pools or in symplectitic intergrowths.

Plagioclase is the dominant phase, varying very little in composition, the average anorthite content being 24 per cent. It occurs with an interlocking, polygonal granoblastic habit, equilibrium apparently being undisturbed apart from a little granulation. Albite twins are well developed in most grains, deformation twinning being rare, and some grains are untwinned. Aligned pools of exsolved quartz are quite common.

Pale brown, subhedral anthophyllite grains, up to 5 mm long, define a strong linear element in the rock as elongate aggregates. They often contain intergrowths and pools of quartz suggesting their growth from an earlier ferromagnesian phase, possibly orthopyroxene. Brown biotite-quartz symplectite replaces anthophyllite in places. Anthophyllite is only found as a major phase in metamorphic rocks when certain chemical conditions are fulfilled by the host rock. The rock must be calcium deficient, or sufficient aluminium must be present to allow all available calcium to be contained in the plagioclase felspar stable under the prevailing metamorphic conditions. Conversely, the rock must be rich in magnesium relative to calcium. If these conditions are not present, then calciferous amphiboles would form in preference to anthophyllite. Rocks fulfilling the conditions could be ultrabasic in composition, or aluminous metasediments. The low anorthite content of the felspar in the anthophyllite-gneisses suggests that aluminium is sufficient relative to calcium to allow calcium to be held in stable plagioclase. Anthophyllite-gneisses are common in high grade metasediment belts throughout world Pre-Cambrian basement rocks, (Rabbitt 1948), and their usual content of cordierite and kyanite suggest that they are aluminous metapelites.

Accessory minerals include apatite, ore and garnet. The mineral assemblage represented by anthophyllite-gneiss is:

plagioclase-anthophyllite-biotite.

Quartzites.

Impure quartzites are important members of the metasediments of Coll and Tiree. They are frequently found as ribs up to a metre thick amongst the other metasediments. Their largest area of outcrop is on Ben Feall, where a 400 m wide band of variegated quartzites, with interbanded calc-silicate rocks, is found (Map 1). These quartzites are white, pink and green, containing numerous small grains of apple green epidote, pale green diopside and dark green tremolite-actinolite. The banding is not mineralogical, but caused by the colour of quartz grains, and it possibly represents original sedimentary layering. Some of the thinner bands contain pyrites, chlorite, graphite and a few are of pure quartz. Where quartzites abut the gneisses, they are cut by pink quartz-felspar pegmatoid veins and show progressive growth of pink felspar towards the margin, presumably a metasomatic effect involving alkali and aluminium diffusion.

Quartz is usually amoeboid with sutured margins and strongly polygonised, although a more equilibrated habit can be seen between the larger meandrine grains, and in some cases quartzites can be a 0.5 mm mozaic of slightly interlobate polygons.

Microcline is common as interstitial pools and amongst polygonal quartz grains where it has recrystallised.

Plagioclase of andesine composition is sparsely developed, and usually saussuritised.

Diopside occurs as skeletal corroded grains up to 1 mm in diameter, usually seen to be replaced by large sieved tremolite grains.

Tremolite derived by hydroxylation from pre-existing diopside occurs as skeletal grains up to 1 cm long, full of quartz blobs.

A planar fabric is seen in some quartzites, defined by pale mica flakes.

Epidote is ubiquitously developed as 0.5 mm grains in bands, and often replacing tremolite.

Accessory minerals include calcite, kyanite, and pyrites.

Like the other rocks, quartzites show evidence of the hydration and retrogression of an early anhydrous mineral assemblage, and the quartzes and felspars show the effects of deformation and subsequent recrystallisation.

These quartzites are probably the metamorphosed equivalents of psammitic

sediments of varying purity, ranging from pure quartzite to calcareous felspathic quartzite.

Calc-silicate rocks.

Pale green, carious weathering calc-silicate rocks, with the silicates often projecting from a carbonate-rich matrix are found as bands within the metasedimentary belts. The quartzites of Ben Feall contain numerous banded ribs of calc-silicate rocks, and at Sloc na Stur banded calc-silicate rocks contain prominent pods of green diopside-tremolite and purple scapolite. Within the quartzo-felspathic gneisses calc-silicate rocks are found as small pink and green pods, and schlieren in the nebulitic gneisses (Plate 13c). They are sometimes zoned, having a garnetiferous core surrounded by green diopside-rich granular material. Banding in these pods may display relic folds.

Textures are largely granoblastic, and grain size ranges from 0.5 mm to 2 cm. The finer grained samples are usually from small pods, the coarsest from thick bands in the metasediment belts.

Quartz is largely absent as a primary mineral.

Microcline occurs in interstitial pools in most rocks but can be very prominent.

Plagioclase, average composition An_{38}, is the dominant mineral. It is largely altered, often completely. However, the outlines of altered grains frequently show equilibrated textures, often being arranged in polygonal aggregates, especially within the pods and schlieren in the migmatites. This is strong evidence for complete recrystallisation of the dislocated pods after their tectonic emplacement in the gneisses. Scapolite, often as large polygonal crystals, is the commonest mineral replacing plagioclase with associated epidote and clinozoisite.

Calcite, as a primary mineral, is variable in amount, but common in late veins with epidote and prehnite..

Diopside grains, 0.5 mm to 2 cm across, are ubiquitous in these rocks, often skeletal. Diopside has been replaced by pale green tremolite, either as intergrown epitaxic grains, or as fibrous mats. Tremolite has recrystallised as polygonal aggregates in some rocks.

Orange garnets are only found in the cores of some pods in the quartzo-

felspathic gneisses, where they occur as amoeboid rims to saussuritised plagioclase and pale green diopside in contact with quartz. Schlieren of plagioclase, diopside and garnet are enclosed in a matrix of elongate quartzes, irregular zoisite plates occurring along quartz boundaries (Plate 17a). This would suggest that garnet has formed by reaction between diopside and plagioclase, excess calcium and aluminium ions contributed to an intergranular fluid, reacting with quartz in the matrix to form zoisite. The reactions could be summarised:

$$\text{Diopside} + \text{plagioclase} \rightleftharpoons \text{grossular} + \text{quartz} + Ca^{2+} + Al^{3+},$$

$$\text{Quartz} + Ca^{2+} + Al^{3+} + H_2O \rightleftharpoons \text{zoisite}.$$

The reaction has gone to completion in some schlieren which consist of meandrine garnet-quartz intergrowth and clouded plagioclase. The anhydrous reaction

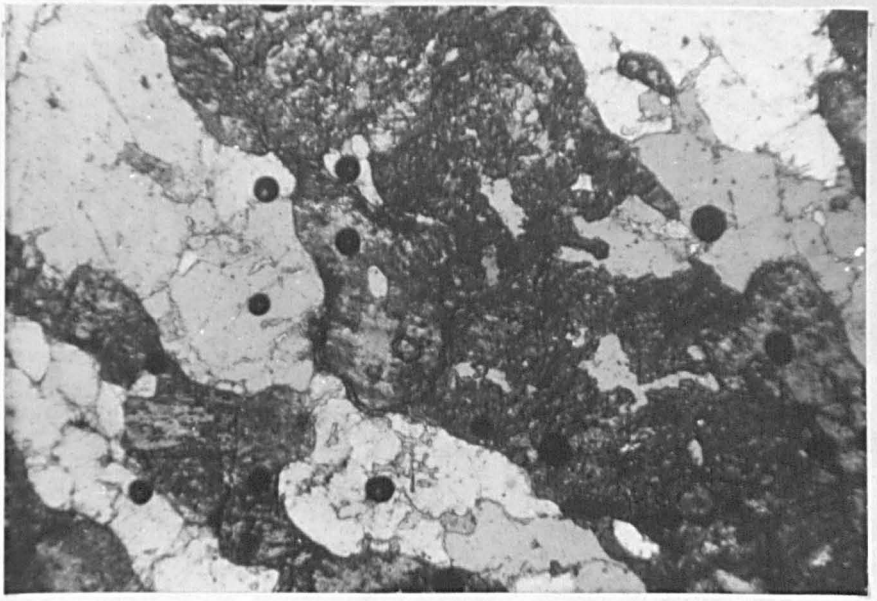
$$\text{plagioclase} + \text{diopside} + \text{ore} \rightleftharpoons \text{garnet} + \text{quartz},$$

suggested by Griffin and Heier (1969) could have taken place, all the ore having been used, leaving the related minerals in equilibrium. This would be supported by the presence of ore in the outer diopside-rich, garnet-free zone, and its absence from the garnetiferous cores. It would only be conclusive of garnet rimmed ore in the cores, and if electron probe analyses showed that the garnets were more iron rich than the pyroxenes (Griffin and Heier 1969 p. 103). The presence of zoisite in the quartz matrix and as rims to some diopside and plagioclase would suggest that the reaction was more complex and involved the H_2O molecule. The texture represents a retrogressive reaction. A similar reaction is postulated for growths of garnet up to 10 cm across in some calc-silicate bands. Pale biotite and phlogopite are common. Prehnite may form the bulk of some calc-silicate rocks both as well formed grains and as fibrous masses.

Accessory minerals include apatite, ore, and sphene rimming the ore.

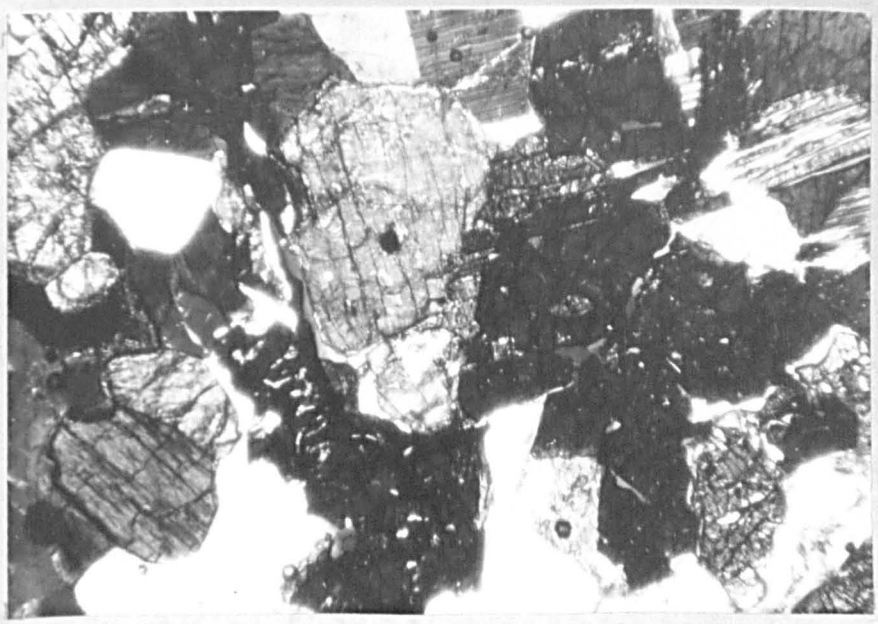
Marbles.

Rocks composed dominantly of calcite or dolomite are present on the islands, being noted for their ornamental qualities. Quarries at Balephetrish on Tiree contain pink fine grained marbles with diopside grains, and white crystalline marbles with green, yellow and orange serpentine spots. These marbles are enclaves within the large metabasite mass, either xenoliths or tectonic selvedges. On Coll 20 m lenses of marbles occur on Soa and near



a

Streaks of clinopyroxene and altered plagioclase with skeletal rims of garnet, separated by quartz with intergranular zoisite grains, in calc-silicate pod from gneiss.
(Spec. 5261d, p.p.l., x 60)



b

Orthopyroxene mantled by clinopyroxene, separated by narrow quartz film from irregular garnet bottom right. Garnet stringer cuts across brown hornblende top left, mafic granofels.
(Spec. 64SW3A, x.p., x 60)

Uig in the Hogh-Breachacha metasediment belt, and at Gorton in the Clabhach-Gorton belt.

In thin section these rocks consist of finely granulated (0.2 mm) calcite sometimes recrystallised, set with larger irregular calcite grains, serpentinised forsterite, diopside, tremolite, scaptolite, phlogopite, chondrodite and clinohumite.

Other metasediments.

Rocks of schistose nature are rare and confined to thin bands up to 5 m thick within other metasediments. The most common is pink and black biotite-schist with prominent beaded quartzo-felspathic bands. Muscovite-schists are sometimes associated with quartzo-felspathic bands in the metasediments. Graphite-schists of undoubted sedimentary origin occur at only two localities, Hogh Bay on Coll (165572) and Scarinish Pier on Tiree (048456). A magnetite-clinopyroxene band occurs with garnet-biotite-gneisses between Bhasapol and Cean a Mhara, it sometimes carries garnets. An enigmatic quartz-plagioclase-garnet-brown hornblende-leucoxene rock is associated with graphite-schist at Hogh Bay, the leucoxene forms the bulk of the rock in places.

Discussion.

The garnet-biotite-gneisses appear to have been most susceptible to metamorphic changes, and a listing of their mineralogical changes will illustrate the metamorphic history of the metasediments.

1) Deformed garnets include sillimanite in helycitic trails. Sillimanite is stable at high temperatures and moderate pressures inverting to kyanite with increase in load pressure.

2) Growth of orthopyroxene and kyanite at granulite facies. Recrystallisation of felsic minerals.

3) Possible reaction of orthopyroxene to form sieved garnets.

4) Partial hydroxylation of orthopyroxene to form blue-green amphibole rims. Deformation of felsic minerals and crystallisation of biotite along new foliation.

5) Nucleation and static growth of garnets. Crystallisation of anthophyllite, cummingtonite or green hornblende as a result of hydroxylation

and breakdown of orthopyroxene, possibly via the stage 4 blue-green amphibole rims.

6) Partial replacement of garnet by biotite, and amphiboles by biotite/quartz symplectite.

There seems to have been no sharp dividing line between the successive stages, and reactions are incomplete in many specimens. For instance stage 3 probably continued through stage 4. The successive events are in accord with the retrogressive metamorphism seen in the quartzo-felspathic gneisses. However, the presence of sillimanite within deformed garnets in a foliation partially defined by kyanite indicates metamorphism prior to granulite facies. Early sillimanite might have formed at upper amphibolite facies, reacting with biotite in a dehydration reaction at granulite facies to give garnet and alkali felspar (De Waard 1965, p. 167). Kyanite crystallised under conditions of higher pressure either during the granulite facies, or at the onset of deformation after the granulite facies.

The mineral assemblages of the other metasediments are limited in their variation by the chemical composition of the rocks. However, they show a similar progression from anhydrous to hydrous assemblages. The textures indicate deformation of equilibrium parageneses, culminating in a second recrystallisation. The growth of garnet in calc-silicate pods indicates amphibolite facies conditions associated with the latter recrystallisation.

In summary, the features of interest shown by the metasediments are:

a) The metasediments represent a varied series of well differentiated sediments, in existence before the granulite facies.

b) There is evidence of a lower grade metamorphic event before the granulite facies in aluminous garnet-biotite-gneisses.

Granofelses.

The hypersthene-gneisses of Ben Hynish are cut by thin concordant buff granofelses less than one metre thick, which give off thinner discordant apophyses (Plate 19a). Such rocks are present as concordant bands and boudins in the adjacent garnet-biotite-gneisses at Hynish. In the latter case the granofelses show faint metamorphic banding, pyroxene-rich bands alternating with bands containing garnet and hornblende. In some, garnet is accompanied by much biotite.

Large areas of garnet-free mafic granofels occur between Ben Hough and the west coast on Tiree, where no foliation is apparent except that defined by sparse concordant white garnet-granites and coarse bands of garnet diopside and hornblende.

Usually, granofels belts are closely associated with metasediments, granofels separating metasediments from quartzo-felspathic gneisses. Garnet-biotite-gneiss with a magnetite band, running from Bhasapol to Ceann a' Mhara on Tiree is bounded on its eastern margin by mafic granofelses, some garnetiferous, others containing pyroxenes and hornblende. A wide granofels horizon follows the western margin of the Hogh Bay - Breachacha metasediments and is folded by a large F_2 fold at Port na Luing. A similar band outcrops along the west side of Crossapol Bay (Map 1). The prominent granofels bands on Coll contain red garnets up to a centimetre across with finer biotite, bluish felspar, opalescent quartz, and sometimes black hornblende substituting for garnet. Garnet-free, hornblende-biotite-granofelses are more foliated than the garnet bearing types which are in complete contrast to the rest of the complex, being devoid of any macroscopic facric.

Granofelses may outcrop as thick belts and discontinuous bands in the migmatitic gneisses, without associated rocks of undoubted sedimentary origin. Betweeen Sorisdale and Arinagour Pier on Coll, a belt some 500 m across at its widest, petring out to the SW, is associated with a similarly thick belt of metabasic rocks, within the migmatitic gneisses. It is identical to those granofelses associated with metasediments. Thinner granofels bands of a discontinuous nature are widespread in the gneisses. Some are present in the nebulitic gneisses of Rudha Hogh, often having more of a foliation, and quartz and felspar porphyoblasts, but carry garnets, are devoid of gneissose banding and recognisable as distinct from the more quartzo-felspathic metasedimentary garnet-biotite-gneisses.

Diopside-granofelses are found in association with diopside-gneisses. Thin boudinaged bands and enclaves are found in diopside-gneiss, and are devoid of garnets but contain prominent green clinopyroxene as well as biotite. Such granofelses maintain their fabric-free appearance in this location.

Where affected by late shearing and mylonitisation granofelses become greyish-green due to the presence of chlorite.

The fabric-free nature of the granofelses is well illustrated by Plates 3c and d.

Mafic granofelses of Tiree.

These rocks display a fabric-free granoblastic texture marked by the presence of straight grain boundaries and triple junctions, especially amongst the felspars. Grain size is about 1 mm.

Quartz only occurs in minor quantities and primary alkali felspar is absent in the mafic granofelses.

Plagioclase ranging from An_{40} to An_{58} is co-dominant with mafic minerals. It can be untwinned or show varying development of polysynthetic deformation twins. It is unzoned and antiperthitic.

Slightly pleochroic orthopyroxene, sometimes showing equilibrium relations to plagioclase, clinopyroxene and brown hornblende, is common. Some grains show undulose oblique extinction and fine polysynthetic twins indicative of straining. Larger grains are irregular and skeletal, embaying plagioclase and partially enclosing polygonal aggregates of felspar. Slight alteration of orthopyroxene to pale yellow serpentine was succeeded by the growth of a mantle of blue-green amphibole.

Clinopyroxene, as well as forming discrete polygonal grains, rims orthopyroxene and ore, occurs with interstitial quartz between ore, garnet and orthopyroxene (Plate 17b), and is contained with quartz in sieved garnets. Seemingly two generations of clinopyroxene are present, one of primary origin, the other associated with a later reaction.

Subhedral, highly pleochroic brown hornblende is equilibrated with both pyroxenes and plagioclase. It is frequently in bands separated by hornblende-free layers, which indicates some variation in water fugacity at granulite facies. Brown metamorphic hornblendes owe their colour to a high titanium content (Leake 1969) and are typical of hydrous sub-facies of the granulite facies (De Waard 1965). Analysed hornblende 78NW1B from a mafic granofels has such a high TiO_2 content (2.08% Table 2). Retrogression has given olive green hornblende full of dusty ore, some of which has coalesced to ilmenite grains. A similar green, but clear, hornblende shows epitaxic replacive textures to clinopyroxene.

Pale pink garnet shows distinctive disequilibrium textures, and its growth was later than the assemblage orthopyroxene-clinopyroxene-brown hornblende-plagioclase. Amoeboid grains up to 2 mm in extent include blobs of quartz and intergrown clinopyroxenes in optical continuity. Also included

in garnet are corroded orthopyroxene grains, and where garnet abuts external orthopyroxene grains there are thin mantles of clinopyroxene (Plate 17b). When in contact with plagioclase, garnets are thinly mantled by quartz (Plate 17b). From such evidence it is clear that a garnet producing reaction involved the phases orthopyroxene, plagioclase, ore, clinopyroxene, garnet and quartz (Plate 17b). Brown hornblende, which is cut by garnet stringers, seems to have been uninvolved, and the reaction was probably anhydrous and initiated the disappearance of orthopyroxene. Some recrystallisation or discrete nucleation of garnet is shown by sparse unsieved euhedral grains up to 0.5 mm across. Large subhedral garnets show marked concentration next to garnet-biotite-gneiss, concordant granitic veins and in coarse bands of garnet and clinopyroxene which are well displayed at Eilean Ghreasamuil. Green hornblende replaces clinopyroxene in such bands.

No primary biotite is present in the mafic granofelses but a secondary, foxy red variety replaces garnet and to some extent orthopyroxene as a complex symplectitic intergrowth with quartz (Plate 18a). The intergrowths are up to 3 mm long and have partly recrystallised to biotite laths with surrounding quartz lenses. Strongly coloured biotites are common in granulite facies rocks and are usually titaniferous (Engel and Engel 1960).

Accessory minerals are apatite, ore, zircon and rutile.

Stable assemblages represented, including quartz and plagioclase, are:

orthopyroxene-clinopyroxene

orthopyroxene-clinopyroxene-brown hornblende,

and metastable:

orthopyroxene-clinopyroxene-garnet-brown hornblende.

Garnet-biotite-granofelses.

These granofelses are equigranular granoblastic and largely devoid of penetrative fabric. Equimensional mineral interfaces are generally interlobate but straight boundaries and triple junctions are sometimes seen. The grain size is between 0.5 and 1 mm.

Quartz is present from 1 to 10 per cent, large grains showing varying degrees of polygonisation.

Plagioclase felspar is the dominant felsic mineral ranging in composition from An_{30} to An_{45}, the most calcic being in orthopyroxene-bearing varieties of granofels. It is usually fresh and saussuritisation is variable.

a

Biotite/quartz symplectite replacing garnet in mafic granofels.
(Spec. 64SW3A, p.p.l., x 60)

b

Randomly oriented biotites in garnet-biotite-granofels. Note the presence of irregular sieved garnet, green hornblende and partially serpentinised orthopyroxene relics, being replaced by biotite.
(Spec. T5A, p.p.l., x 60)

Some grains are untwinned but regular albite twins crossed by thinner pericline twins are usual. Deformation twins are largely absent and the regularity may be the result of recrystallisation or coalescence of spindle twins. Zoning with sodic margins is common.

Orthopyroxene is present in a few specimens as small (less than 0.5 mm) grains of irregular embayed or rounded forms. They are cracked and partly serpentinised, an unbroken thin rim of blue-green amphibole surrounding them. Streaks of granulated orthopyroxene follow a planar fabric when present. There is marginal alteration to small (0.1 mm) intergrowths of biotite and quartz (Plate 18b).

Quartz-sieved, green hornblende is in sparse amounts and replaces orthopyroxene and cummingtonite, itself possibly replacive to pre-existing orthopyroxene. Green hornblende often mantles garnet and is enclosed by it. Whether it replaces garnet directly, or not, is impossible to determine, but retrogression and hydration would be expected to favour the breakdown of garnet and the growth of hornblende.

Pink garnet is plentiful in these rocks, in two distinct habits. Fairly large (2 mm) garnets of irregular to amoeboid outline are full of quartz inclusions and may contain orthopyroxene, hornblende and biotite (Plate 18b). They rarely show helycitic trails. Such textures suggest their origin as products of early reactions, perhaps similar to that demonstrable in mafic granofelses. They sometimes have inclusion free rims which may represent marginal clearing of inclusions or later superimposed growth. Smaller (0.2 mm) euhedral unsieved garnets are sited within other minerals and on grain boundaries. Those on grain boundaries are several orders of magnitude larger than those within felspars. These smaller garnets probably grew at independant nucleation sites, the range in size indicating the part played by grain boundary diffusion during their growth. They are often associated with clusters of biotite whose cleavage they truncate and may have been partly derived from earlier biotites' changing composition in response to nearby garnet growth. The inclusion free rims of the large garnets are preferentially developed in biotite-rich environments. Some of the large irregular garnets are replaced by penetrating biotite flakes and tangentially arranged mantles of biotite.

Randomly oriented biotite is the dominant mafic phase, rarely being roughly aligned (Plate 18b). It is a strongly pleochroic, foxy red variety rich in TiO_2 (Table2, 3.3%). As has been described it is of several genera-

tions. Because of its flexible composition biotite may have changed its chemistry in response to the appearence and disappearence of other ferromagnesian phases, yet still remains as a stable phase. Rast (1965) has suggested that phylloblastic minerals on recrystallising do not tend to form polygonal aggregates but are mimetically regenerated at the sites of earlier micas. This recrystallisation involves nucleation of new mica of different composition where a favourable lattice is situated. All the biotites in the thin sections examined have the same colour, although they differ in their origins, and it is suggested that all are of a similar composition as a result of ionic exchange with the other stable ferromagnesian phase; garnet.

Accessories include apatite, zircon, ore and rutile.

Stable assemblages, including quartz and plagioclase, are:

garnet-biotite-orthopyroxene

garnet-biotite-hornblende

garnet-biotite.

Hornblende-biotite-granofelses.

These rocks have essentially the same textures as those carrying garnet. They contain prominent sieved green hornblende, yellow-brown biotite, sphene and are rather richer in ore than garnet-biotite-granofels. The biotites are full of small dusty ore grains and are frequently replaced by chlorite and prehnite wedges, then being greenish. Sphene occurs as small rounded grains and rims grains of ore. Epidote is present in some specimens.

The assemblage is hornblende-biotite-plagioclase-quartz.

Hornblende-biotite-granofelses are probably retrogressed garnet-biotite-granofels, garnet having been replaced by hornblende and biotite. Biotites are probably chemically different to those of the garnetiferous varieties, this being reflected by the colour difference.

Diopside-granofelses.

In aspect these are similar to hornblende-biotite-granofelses, but are contrasted by their content of poikiloblastic clinopyroxene grains similar to those in diopside-gneiss. The clinopyroxene has generally been replaced by epitaxically intergrown pale green amphibole and quartz. Like

the diopside-gneisses, which contain such granofelses, they may represent a metasomatic development in favourable sites.

Discussion.

The lack of foliation and segregation banding in the granofelses, which are surrounded by foliated rocks, and whose content of isoclinally folded concordant granitic rocks shows that they have been deformed, is their constant and enigmatic feature. The content of biotite in the commonest varieties would not preclude the formation of a tectonic fabric, but in thin section biotites are randomly oriented in general. Originally banded rocks would tend to have planar inhomogeneities emphasised in metamorphism by ionic diffusion caused mainly by compositional gradients between adjacent bands. Relatively homogeneous Lewisian basic rock at Gairloch developed segregation banding by this means during amphibolite facies metamorphism (Bowes and Park 1966) as have metabasites on Coll and Tiree. To be still lacking in banding, an origin as homogeneous rocks must be postulated for the granofelses. Few sedimentary rocks of the required intermediate composition (greywackes) lack banding, and the origin of the granofelses of Coll and Tiree by igneous processes is a reasonable possibility. This hypothesis is supported by the presence of discordant mafic granofelses in hypersthene-gneisses on Tiree. It is discussed further in Chapter 4. Another possible mode of origin, excluding the discordant mafic granofelses, is as mylonite bands. Mylonites although usually colour banded are frequently very homogeneous fine grained rocks and sometimes glassy.(Christie 1960). Large amounts of mylonites are present in many metamorphic terranes such as the Moine Thrust Zone, and as late stage bands in the Loch Maree Lewisian sediments.

Granofelses were in position before granulite facies metamorphism, and incompletely retrogressed examples contain the assemblages orthopyroxene-clinopyroxene and orthopyroxene-clinopyroxene-brown hornblende with plagioclase and quartz. This metamorphism did not completely dehydrate these rocks as evidenced by the irregular distribution of brown hornblende.

The earliest mineral transformation after granulite facies parageneses were developed can be best expressed by the equation; orthopyroxene + plagioclase + ore $\rightleftharpoons$ garnet + clinopyroxene + quartz, which was anhydrous.

As in other rock types the onset of hydroxylation of orthopyroxene is marked by the presence of "uralitic" blue-green amphibole rims. Such rims, which might have been akin to cummingtonite, perhaps provided an intermediary phase for alteration of orthopyroxene. Cummingtonites in some garnet-biotite-granofelses are replaced by green hornblende. Other early hydration reactions were:

brown titaniferous hornblende $\rightleftharpoons$ green hornblende + ilmenite,

orthopyroxene + potash felspar component + $H_2O \rightleftharpoons$ biotite + quartz,

clinopyroxene + plagioclase + water $\rightleftharpoons$ green hornblende + quartz.

At this early stage of amphibolite facies retrogression garnet growth at discrete nuclei was accomplished and the typical assemblages were those found now in garnet-biotite-granofelses.

Further retrogression to a lower part of the amphibolite facies saw the disappearence of garnet by

garnet + potash felspar component + $H_2O \rightleftharpoons$ biotite + quartz,

and possible replacement by green hornblende, giving hornblende-biotite-granofelses. The event marked by the formation of nebulitic-gneisses and diopside-gneiss manifested itself on some granofelses by the growth of clinopyroxene poikiloblasts, themselves hydroxylated forming actinolite-quartz symplectites.

The presence of chlorite, prehnite and epidote in granofels is the mark of late greenschist facies metamorphism.

The main points of interest shown by the granofels are:

a) Their present lack of fabric suggests an original lack of planar features, possibly as a result of forming by igneous processes before the granulite facies.

b) An interesting garnet-forming reaction marked the close of granulite facies metamorphism.

c) At the same time as the formation of diopside-gneisses, diopside poikiloblasts grew in associated granofelses.

Metabasites.

Dark coloured, granulitic to banded and foliated metabasites are widespread on Coll and Tiree. They occur as belts up to 500 m thick, in thin bands less than 10 m thick, and as pods and schlieren in quartzo-felspathic gneisses. Their variation is limited by their restricted mineralogy, dom-

inantly hornblende, and several types of different history may be present without distinguishing features in the field.

Metabasites are generally concordant with the prominent foliation. At outcrop, the thinner metabasites may be distinguished from the dark bands in quartzo-felspathic gneiss by their strong contrast to the gneiss, sharp margins and irregular distribution.

In western Tiree, especially between Balephuil and Hynish Bay, concordant metabasites frequently emit thin discordant stringers which cut the foliation and garnetiferous granite bodies in the hypersthene-gneisses (Plates 19b and c). These apophyses are often at high angles to the foliation and displace it in places. Relic igneous features such as phenocrysts, marginal chilling or banding are absent and they are identical in all respects to the metabasites from which they emerge. In massive hypersthene-gneisses on Ceann a' Mhara thin discordant metabasites (Plate 19b), have been demonstrably deformed and transposed into concordant bands and schlieren, the host gneiss becoming progressively more banded (Plate 20a). Prominent bands of grey metabasites up to 40 m thick cut hypersthene-gneisses on Ben Hynish. They are concordant on a wide scale but show slight discordance locally. Garnets and pyroxenes are prominent. Thus it is reasonable to state that in the west of Tiree metabasites exist as dyke-like bodies, and at least some of the concordant bodies have an intrusive origin.

The concordant metabasites in amphibolite facies terranes show varied characteristics at outcrop. Rarely they show slight discordance (Plate 19d). Their fabric varies from being granoblastic, lineated or foliated, and they show different degrees of mineralogical banding. Their foliation varies with the state of deformation in the host rocks, metabasites in hypersthene-gneisses showing no planar fabric. It also appears to be a function of thickness, thick bodies showing stronger foliation and banding than nearby thinner bands in the migmatitic gneisses. The foliation was present before D_2. It is often marked by mineralogical banding (Plate 20b).

Frequently the foliation contains deformed diopside-and garnet-rich specks, whilst the bulk of the rock is hornblendic. These specks pre-date a later portion of the deformation whilst diopside and garnet in diffuse bands in the hornblendic parts of the same outcrop were controlled in their growth by an earlier phase of the deformation encouraging the formation of banding. The process involved in the production of banded metabasites was undoubtedly metamorphic, similar to that suggested by Bowes and Park (1966)

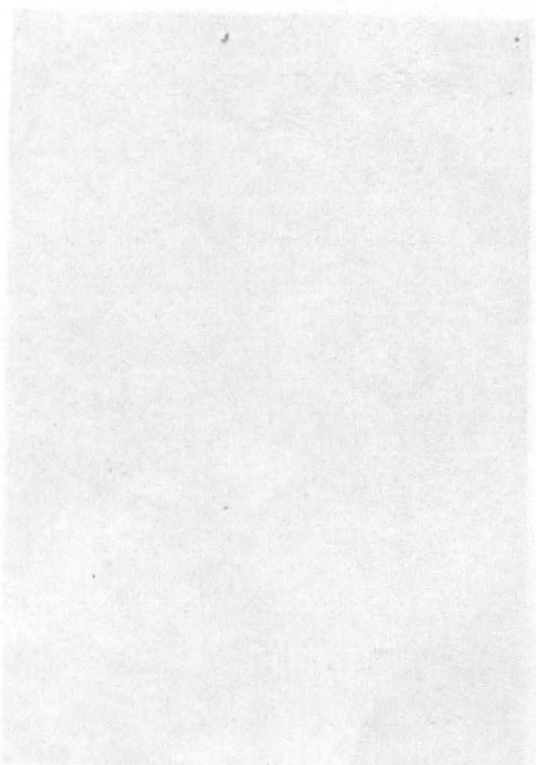


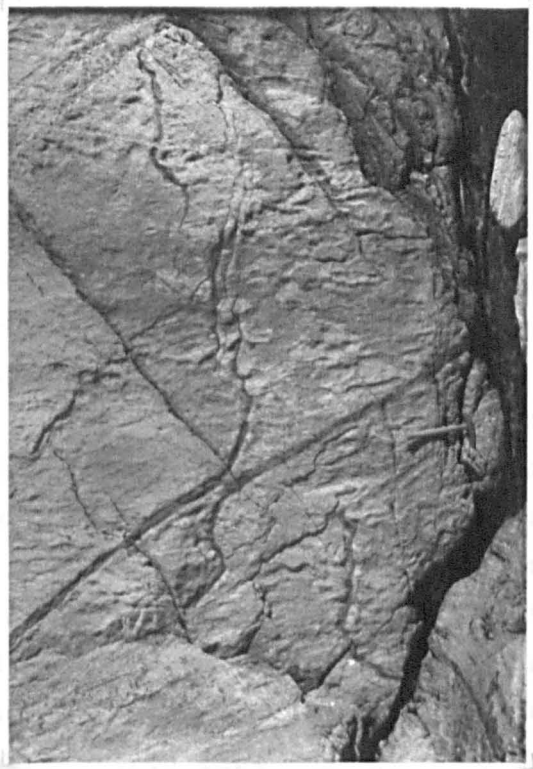
Plate 19

a. Discordant mafic granofels cutting poorly foliated hypersthene-gneiss: Ben Hynish, Tiree.

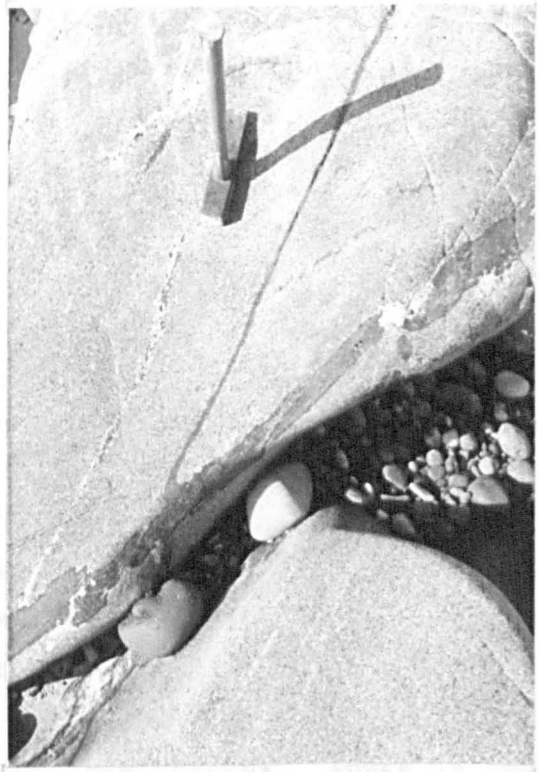
b. Metabasite cutting massive hypersthene-gneiss: Ceann a' Mhara, Tiree.

c. Metabasite bands with discordant apophyses in well foliated hypersthene-gneisses with diffuse speckled garnet- and hornblende-rich bodies: Hynish, Tiree.

d. Discordant metabasite cutting migmatitic gneiss: 243637, Coll.



a



b

c

d

for banded Lewisian basic rocks at Gairloch.

Metabasite bands often 100 m thick frequently margin the metasediment belts, in association with granofelses (Map 1). Whether they are related to the metasediment belts either in origin or because of preferred intrusion, or whether they protected the metasediments from metasomatic alteration to quartzo-felspathic gneiss, will be discussed elsewhere in the text. There is a common association of metabasites with granofels in most occurrences (Map 1).

In the migmatitic gneisses and metasediments, metabasites are generally in relatively thin boudinaged bands. In the nebulitic gneisses they are represented by dislocated enclaves and schlieren.

On Coll a prominent feature of metabasites on all scales is the presence of an agmatitic structure (King, 1965), where angular fragments of metabasite and hornblendite are set in an anastomosing matrix of pale speckled granitic material (Plate 20c). Areas of agmatite range in extent from hundreds of square metres shown on Map 1 to only a few square metres. There is much variation in fragment size and angularity. The felsic material is coarse grained and may have sharp margins or pass through a diffuse speckled hornblendic variety into metabasite. Unlike examples described from the Lewisian of Benbecula by Dunning and Dearnley (1968) and from Drumbeg by Sheraton (1970) the felsic portions rarely display a later foliation. Plate 20d shows rootless areas of granitic material in a metabasite, giving it a spotted appearence, a hornblende-rich rim surrounds each felsic pool. This suggests that the amphibolite itself produced felsic fluids by some process, thereby becoming more mafic. Ultramafic fringes between a metabasite and host gneiss and discordant felsic veins developing in incipient cicatrices in a metabasite band are shown in Plate 10a. Invariably, the surrounding gneisses have a large content of pegmatite veins and the best developed agmatites are found in nebulitic gneiss terranes (Plate 21a) where the process has been carried to extremes as shown by Plate 13d and described in Chapter 2. Agmatisation in this case seems to have been of dominantly tectonic nature, fragmentation of competent metabasite bodies having occurred in strongly deformed migmatitic and nebulitic gneisses. Cicatrices opened by boudinage of metabasites would tend to be filled with the products of anatexis or selective hydrous solution in the surrounding acid gneisses as has been suggested by Ramberg (1956 p. 189). Metasomatic exchange with the basic fragments would add to this, basifying the fragments. The process

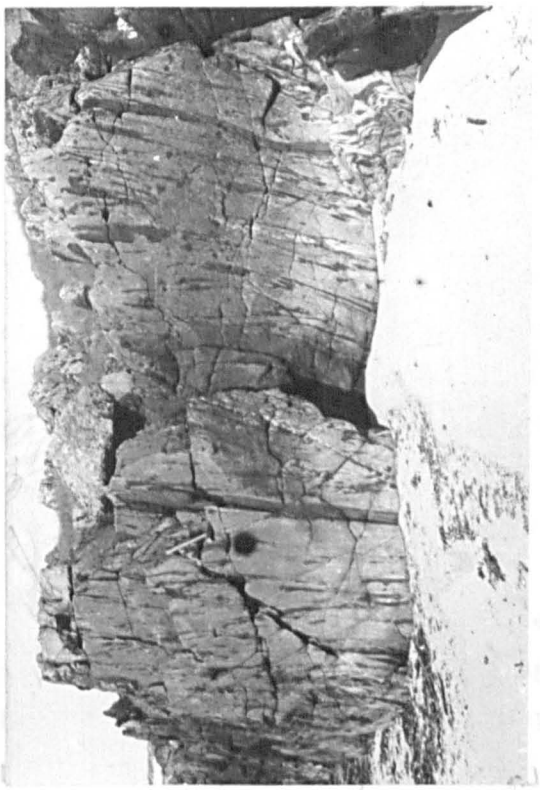
Plate 20

a. Schlieren of metabasites in foliated migmatitic gneiss, some of which can be traced into dykes: Ceann a' Mhara, Tiree.

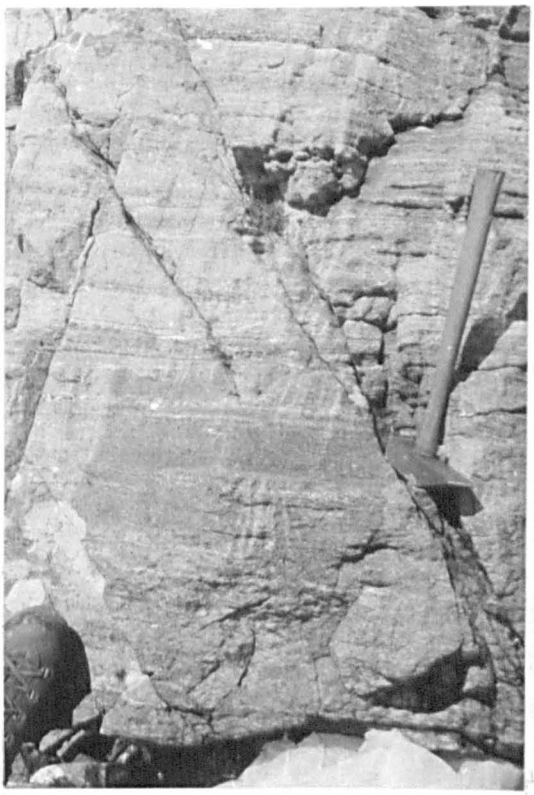
b. Banded garnet-amphibolite: Sorisdale, Coll.

c. Well developed agmatite: Pier road, Coll.

d. Rootless granitic bodies in metabasite, note the presence of hornblendite rims: 220610, Coll.



a



b

c

d

would enable the competent metabasites to be dislocated with the less viscous host rocks. Agmatites of similar appearance have been described from amphibolite facies terranes in many Precambrian Shield areas (Berthelsen 1960; Ramberg 1952, 1956; Heier 1960; Dunning and Dearnley 1968).

Petrographic description of metabasites is subdivided according to the presence or absence of orthopyroxene and garnet in thin section after the scheme proposed by Berthelsen (1960). In mafic metamorphic rocks ferromagnesian mineral content is greater than felsic mineral content.

Pyribolites and pyriclasites.

These terms were coined by Berthelsen (1960) and apply to granoblastic mafic metamorphic rocks containing both ortho- and clinopyroxene. Pyribolites contain a proportion of amphibole between 33 and 66 per cent of the mafic minerals, pyriclasites less than 33 per cent. They are modified by petrographic and mineralogical prefixes.

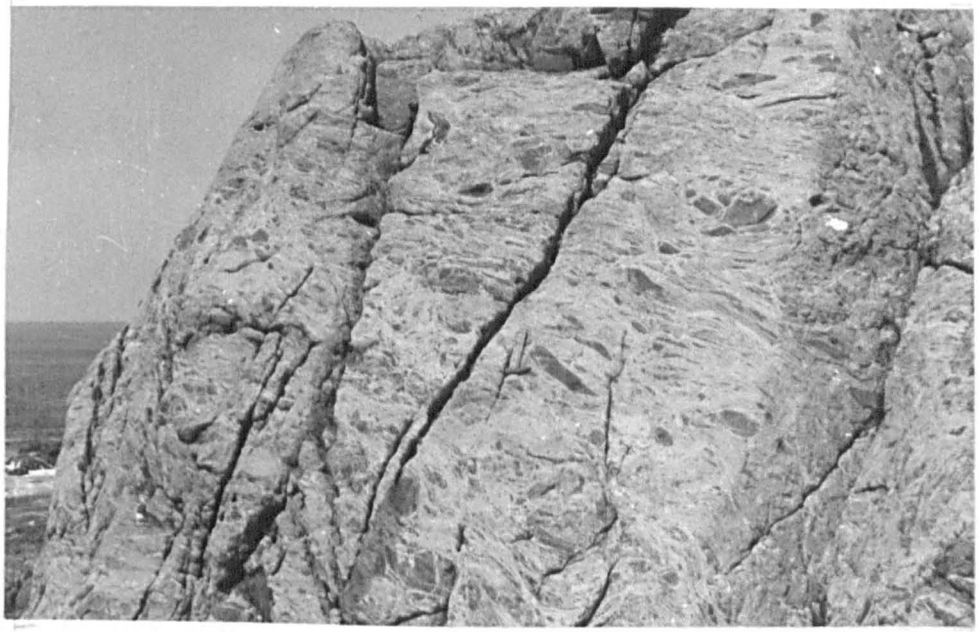
On Coll and Tiree pyribolites and pyriclasites are granoblastic with a grain size up to 2 mm. They are the commonest mafic rock on Tiree and are found in a few localities on Coll noteably at the west side of Hogh Bay (Specimen T1A). They are generally concordant, but the discordant bodies of Tiree are all pyribolites or pyriclasites.

Quartz is present only in minor amounts as interstitial pools and included in other minerals.

Plagioclase of average sodic labradorite composition makes up to 40 per cent of the rock, averaging 30 per cent. It is usually very fresh and occurs in pyriclasites as polygonal grains indicative of its equilibration. In pyribolites the grains are interlobate. Plagioclase from pyriclasites shows regular albite twins similar to those in igneous rocks, indicative of completion of inversion or annealing in the solid state in the granulite facies (Spry 1969 p. 81). In pyribolites spindley strain twins are typical.

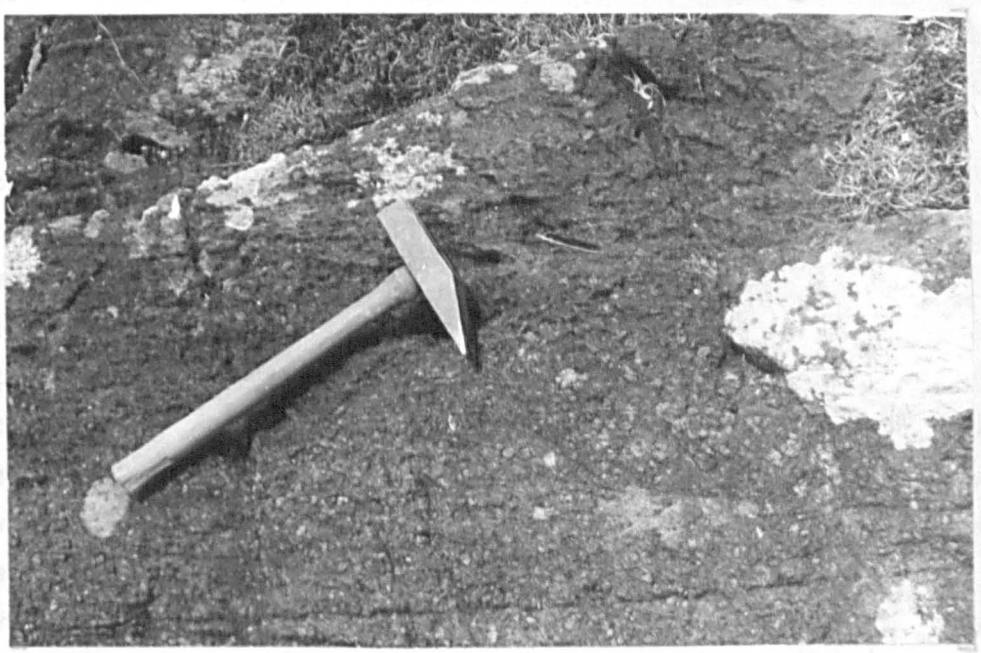
Alkali felspar is not present in these rocks.

Pleochroic (pink to pale green) orthopyroxene is a critical mineral in pyriclasites and pyribolites and is present in amounts up to 5 per cent. It is usually fresh but may be altered to a yellow-green serpentine-like material along cracks. Some orthopyroxenes have a thin mantle of blue-green amphibole. Individual grains of orthopyroxene may be subhedral and



a

Agmatite in nebulitic gneiss of Rudha Hogh. The alignment and flattening of mafic fragments is due to D_3 deformation.



b

Unfoliated ultramafic rock with large orthopyroxene crystals: Bousd, Coll.

interlocked with clinopyroxene and brown hornblende in equilibrium. In some cases, however, interlobate grains are rimmed with irregular mantles of clinopyroxene in some reaction relationship (Plate 22b), especially in proximity to garnets and/or grains of opaque ore. Other irregular grains of orthopyroxene have been replaced by a biotite-quartz symplectite which also mantles nearby ore grains. This relationship may represent the reaction:

orthopyroxene + ore + K-felspar + H_2O $\rightleftharpoons$ biotite + quartz,

alkali felspar component having been derived from plagioclase. Or it may possibly have proceeded via the blue-green amphibole rim:

cummingtonite + K-felspar + ore $\rightleftharpoons$ biotite + quartz.

Pale green clinopyroxene is prominent, usually in greater quantities than orthopyroxene. It sometimes contains narrow exsolution lamellae of orthopyroxene suggesting that limited solid solution was favoured by early physical conditions.

Brown hornblende is variable in its distribution from 1 to 20 per cent. An analysis of a brown hornblende from pyribolite specimen 78SW5B has high TiO_2 and (Na_2O + K_2O) (Table 2). Its textures indicate equilibrium with both pyroxenes. Alteration has resulted in olive green hornblende with dusty ore inclusions and grain boudaries marked by ore. This feature could have resulted from titanium leaving the brown hornblende lattice with iron in response to changing metamorphic conditions. Olive green hornblende also demonstrably replaces clinopyroxene in some specimens, occurring as epitaxic intergrowths with the pyroxene. Evidence for its replacing orthopyroxene is sparse. Thus the growth of green hornblende was partly due to hydration and partly to alteration of earlier brown hornblende.

Pale pink garnet is common in both pyribolites and pyriclasites, occurring as amoeboid or skeletal grains and subhedral grains, both types sieved with quartz inclusions. It is present in quantities up to 5 per cent of the total volume. When isolate and set in felspar, garnets are commonly subhedral and small (0.5 mm), being rimmed by a thin quartz film. In proximity to orthopyroxene it rims ore grains, and is separated from orthopyroxene by quartz and sometimes clinopyroxene films (Plate 22a). Included in garnets are quartzes as irregular blobs in a symplectitic relationship with garnet, and irregular grains of brown hornblende and plagioclase which have been incorporated by skeletal garnet growth. In some rare cases clinopyroxene is intergrown with garnet, amoeboid areas of pyroxene being in

optical continuity. These mineral relationships suggest a reaction forming garnet at the expense of orthopyroxene and ore, the involved phases being orthopyroxene, plagioclase and ore, and garnet clinopyroxene and quartz. This is identical to the situation in the mafic granofelses of Tiree and the same reaction equation adequately describes it.

Accessory minerals are apatite, rutile and ilmenite.

The stable assemblages are:

orthopyroxene-clinopyroxene-plagioclase (-quartz)

and orthopyroxene-clinopyroxene-hornblende-plagioclase (-quartz).

Metastable assemblages are:

orthopyroxene-clinopyroxene-hornblende-garnet-plagioclase (-quartz)

and orthopyroxene-clinopyroxene-hornblende-garnet-biotite-plagioclase (-quartz).

Garnet-amphibolites.

Berthelsen (1960) defines amphibolites as basic rocks with roughly equal proportions of mafic and felsic minerals, and with amphiboles being greater than 66 per cent of the mafic mineral content.

Garnet-amphibolites are dominantly granoblastic, although some may have a part planar, part linear fabric defined by green hornblende in preferred orientation. They vary in grain size from 0.5 to 2 mm.

Quartz is more prominent in garnet-amphibolites than in pyribolites. It is found as interlocking interlobate grains with felspar as well as in intergrowths. It is strongly polygonised.

Plagioclase felspar is often clouded, and ranges in composition from An_{30} to An_{50} averaging about An_{40}. Its interlobate texture and spindle twins evidence deformation. Alteration products include fine grained saussurite and radiating prehnite.

Pale green clinopyroxene in 0.5 to 1 mm grains is present in some rocks, evenly distributed with the other ferromagnesian minerals. In others it also occurs as relatively large grains (up to 5 mm) in eyes, associated with garnet. When in contact with green hornblende, clinopyroxene generally shows embayments and is partially replaced by epitaxically intergrown green hornblende (Plate 22b).

Olive green hornblende is the dominant mineral in these rocks and usually occurs as polygonal grains showing signs of recrystallisation.

Its roughly aligned prisms and heterogeneity of distribution help define a penetrative fabric. Where it replaces clinopyroxene it is usually amoeboid and sieved with quartz, whilst others in the rock have a stronger crystalline form. Whether all the hornblende formed in this fashion, or some is a result of recrystallisation of brown hornblende or alteration of orthopyroxene is not clear from the sections examined. As described, recrystallisation of brown hornblende in higher grade rocks gives olive green hornblende, and in some garnet-amphibolites green hornblende engulfs relics of cummingtonite, and cummingtonite shows a colour change culminating in green hornblende.

Pale pink garnets occur as quartz-sieved porphyroblasts up to 2 cm across, sometimes showing helycitic trails. Such large garnets are present in pale eyes with clinopyroxene (Plate 22b) and may well have originated by the garnet producing reaction noted in pyribolites. Other porphyroblasts show inclusion rich cores and clear margins with a subhedral outline. They are often surrounded by a zone depleted in mafic minerals (Plate 23a). Such garnets possibly underwent a secondary growth by intergranular diffusion of ionic components controlled by the presence of earlier formed sieved garnets. Nucleation would be convenienced by the presence of the garnet lattice at early sites. Nucleation independent of early garnet is shown by euhedral grains of garnet at mineral interfaces and within other minerals. These garnets show a marked variation in size up to 0.5 mm which may be attributed to variation in ease of ionic diffusion, or a slow rate of nucleation coupled with a slow rate of growth.

Accessory minerals are ilmenite, apatite, biotite and sphene.

Stable assemblages are:

hornblende-clinopyroxene-garnet-plagioclase-quartz

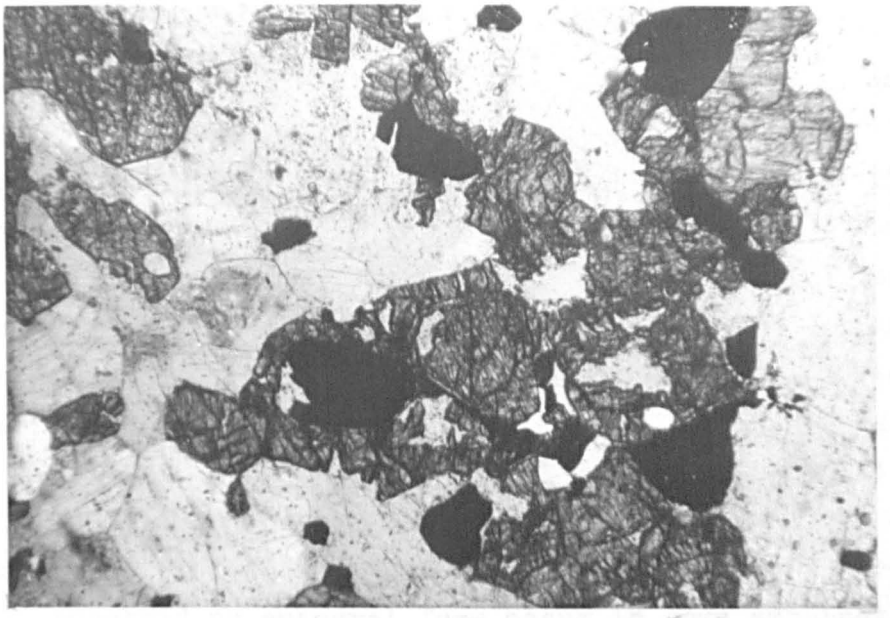
hornblende-garnet-plagioclase-quartz.

Amphibolites.

Mafic rocks composed largely of hornblende and plagioclase are commonest on Coll. They are usually foliated and lineated, grain size varying from 0.5 to 2 mm.

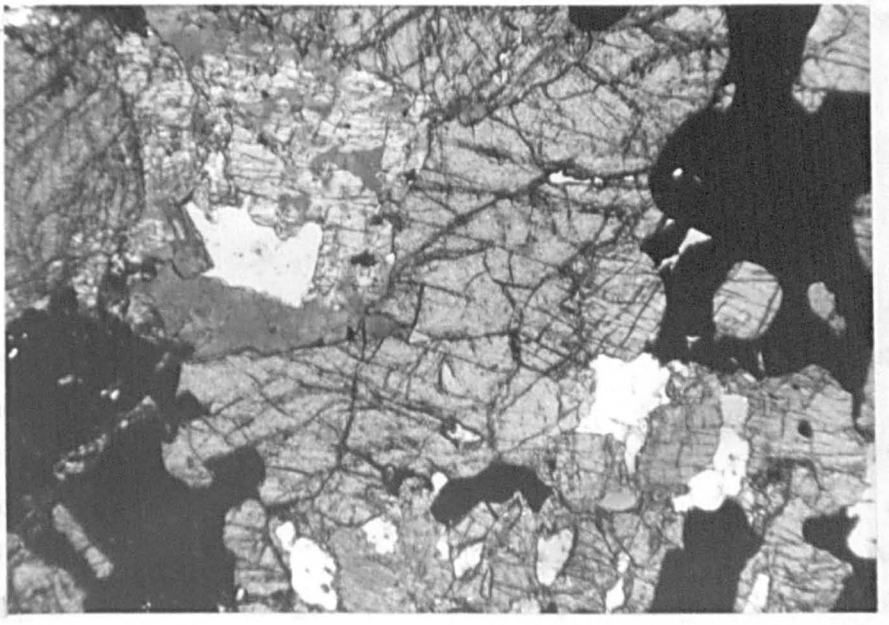
Quartz shows the same features as in garnet-amphibolites.

Plagioclase is often clouded, ranges in composition from An_{20} to An_{35} and shows much evidence of deformation. Alteration products are largely



a

Garnet rimming orthopyroxene and ore, with intervening clinopyroxene and quartz adjacent to orthopyroxene, in pyribolite.
(Spec. 78NE2, p.p.l., x 60)



b

Coarse, partially intergrown garnet and clinopyroxene from an "eye" in garnet-amphibolite. Note the replacement of clinopyroxene by green hornblende.
(Spec. 5781D, p.p.l., x 60)

fine grained but include recognisable sericite, clinozoisite, epidote and pools of clear albite.

Clinopyroxene is present in some sections both in equilibrium with green hornblende and partially replaced by it.

Green hornblende is largely unsieved. In fine grained (0.25 mm) amphibolites it has been granulated and recrystallised as polygonal interlocking grains (Plate 23b). Some hornblende is partly chloritised.

Certain amphibolites show prominent pale spots up to 2 cm across which prove to be composed of fine grained (0.5 mm) saussuritised plagioclase, green hornblende, and much opaque ore in angular grains and along mineral interfaces (Plate 24a). The texture is polygonal and contrasted to that of the rest of the rock and the spots may represent hydrated and altered pre-existing garnets. Orange garnets, as distinct from pale pink ones, with a skeletal, amoeboid or euhedral form occur in some amphibolites from agmatite bodies. They lie in patches of saussuritised felspar and are associated with large pink sphenes. They tend to be extended along relic felspar boundaries as thin stringers. It is suggested that they formed when plagioclase was altered, being a calcium rich variety and taking calcium from the altered felspar.

Accessories are apatite, ore, sphene, biotite, chlorite and prehnite.

Stable assemblages are:

hornblende-plagioclase-quartz

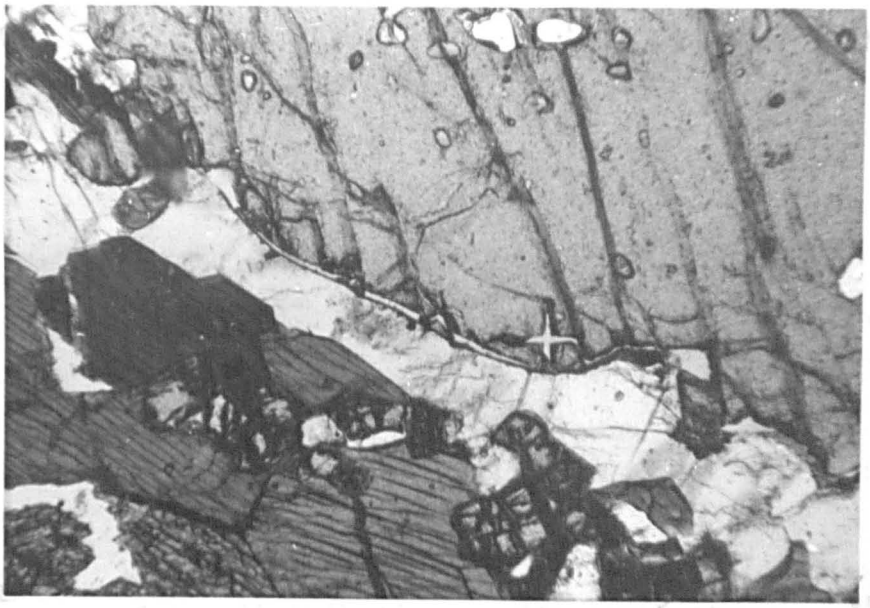
hornblende-clinopyroxene-plagioclase-quartz.

Discussion.

At least some of the metabasites represent intrusive rocks which were subsequently metamorphosed at granulite facies.

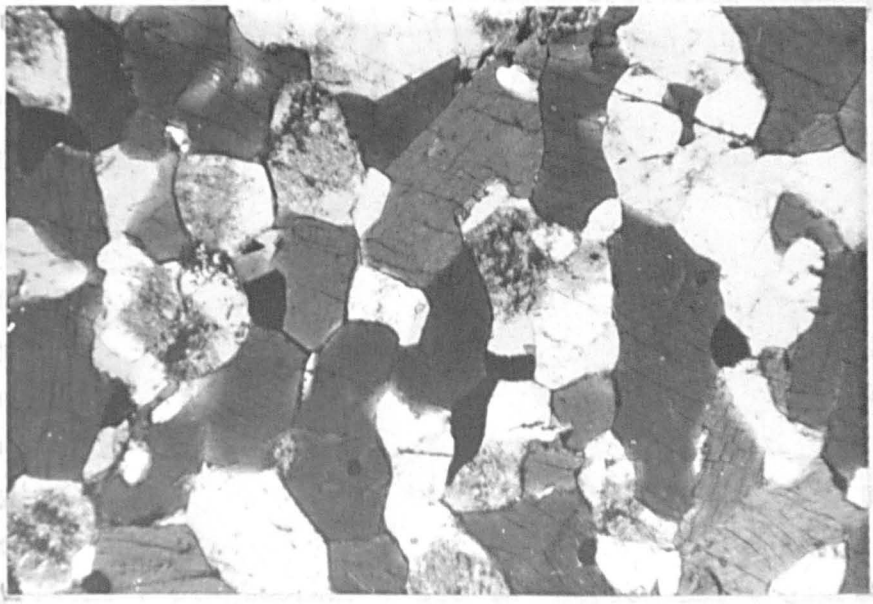
The mineral assemblages present in pyriclasites and pyribolites are indicative of equilibrium crystallisation at granulite facies. According to de Waard (1965) the assemblages plagioclase-orthopyroxene-clinopyroxene (-quartz) and plagioclase-orthopyroxene-clinopyroxene-brown hornblende (-quartz) represent the orthopyroxene-plagioclase- and orthopyroxene-hornblende-plagioclase sub-facies. The irregular distribution of primary brown hornblende in metabasites suggests short range variations in the hydrated state of these rocks at granulite facies.

The first recognisable reaction after granulite facies crystallisation



a

Margin of large sieved garnet, with clear rim in garnet-amphibolite, showing "leached" zone around it consisting largely of quartz and felspar. Small orthopyroxene relics can be seen close to the leached zone.
(Spec. 7345A, p.p.l., x 60)



b

Polygonal hornblende plagioclase and quartz in recrystallised amphibolite.
(Spec. 5781E, p.p.l., x 60)

was anhydrous and may be summarised thus:-

orthopyroxene + plagioclase + ore $\rightleftharpoons$ garnet + clinopyroxene + quartz.

De Waard (1965, 1967) has maintained that the reaction

$$\underset{\text{orthopyroxene}}{6(Fe,Mg)SiO_3} + \underset{\text{anorthite}}{2CaAl_2Si_2O_8} \rightleftharpoons \underset{\text{clinopyroxene}}{Ca(Fe,Mg)Si_2O_6}$$
$$+ \underset{\text{almandine}}{Ca(Fe,Mg)_5Al_4Si_6O_{24}} + \underset{\text{quartz}}{2SiO_2}$$

in metabasites, proceeds to the right with increasing pressure, total molar volume decreasing. He has used it to define high pressure subfacies of the granulite facies.

Saxena (1969) has criticised the validity of this simplification since components of solid solutions are on both sides of the equation. All phases coexist in equilibrium at constant temperature, pressure and water fugacity, and nothing is known of the distribution coefficients of, say, calcium between plagioclase and garnet.

Green and Ringwood (1967) deduced from their experiments on the origin of eclogites that a granulite facies assemblage of orthopyroxene, clinopyroxene and plagioclase should slowly change to an eclogite-like assemblage of garnet, clinopyroxene and plagioclase, if temperature falls at constant pressure after the granulite facies. The transformation would tend to be sluggish. As almandine and clinopyroxene are stable at upper amphibolite facies in basic rocks, the interruption of this static cooling by amphibolite facies metamorphism would result in garnet-clinopyroxene-amphibolites being formed. This was possibly the case on Coll and Tiree where early garnets are overgrown by a second growth, and clinopyroxene is stable with hornblende in garnet-amphibolite. The change from brown, titaniferous hornblende to a green variety might have proceeded during this temperature fall, but no experimental details on the variation of titanium content of amphiboles with temperature are available.

Further hydration and changing physical conditions are shown by the hydroxylation of clinopyroxene to hornblende:

clinopyroxene + plagioclase + $H_2O \rightleftharpoons$ hornblende + quartz

and by the alteration of garnet:

almandine + $H_2O \rightleftharpoons$ hornblende + plagioclase + ore.

Recrystallisation of hornblende resulted in simple green hornblende-plagioclase-quartz rocks. In certain localities the growth of grossular garnet was associated with the alteration of felspar and the growth of sphene.

The presence of chlorite replacing hornblende, and the presence of epidote and albite in altered plagioclase is probably the result of late shearing at greenschist facies. The assemblage albite-actinolite-epidote-chlorite-sphene-quartz is common in greenschist facies basic rocks (Turner and Verhoogen 1960). The absence of actinolite from these Lewisian metabasites is probably due to the metastable presence of hornblende.

The points of interest arising from petrographic study of the metabasites are:

a) Some metabasites occur as pre-granulite facies dykes and are therefore igneous.

b) Initial granulite facies assemblages are consistent with a low to intermediate pressure subfacies of variable dehydration.

c) An anhydrous reaction resulting in the growth of garnet and clinopyroxene may indicate that either pressure increased towards the close of the granulite facies, or pressure was maintained as temperature fell.

d) Complete recrystallisation during the amphibolite facies gave hydrous assemblages which were stable.

e) At least some of the garnet-amphibolites represent a higher grade during the amphibolite facies than amphibolites, as garnets can be demonstrated to have been changed to hydrous minerals. There may be some chemical difference between some garnet-amphibolites and amphibolites.

f) The process of agmatisation was initiated during D_3 in proximity to nebulitic gneisses. It continued through D_4.

Ultramafic rocks.

Rocks in which ferromagnesian minerals are overwhelmingly dominant are of two generations, those originating from ultrabasic bodies and those developed by the process of agmatisation.

Primary ultramafic rocks occur as large mappable bodies at only three localities, Rubha Mor and Clabhach on Coll and between Loch a' Phuill and Ceann a Mhara on Tiree, where they are in the form of zoned, boudinaged bands. The largest, at Rubha Mor, is made up of a series of ovate bodies up to 20 m across set in migmatitic hornblende-biotite-gneisses (Fig. 10). The cores are coarse grained massive dark grey rocks containing prominent pink orthopyroxene crystals (Plate 21b). A rim of black hornblendite is

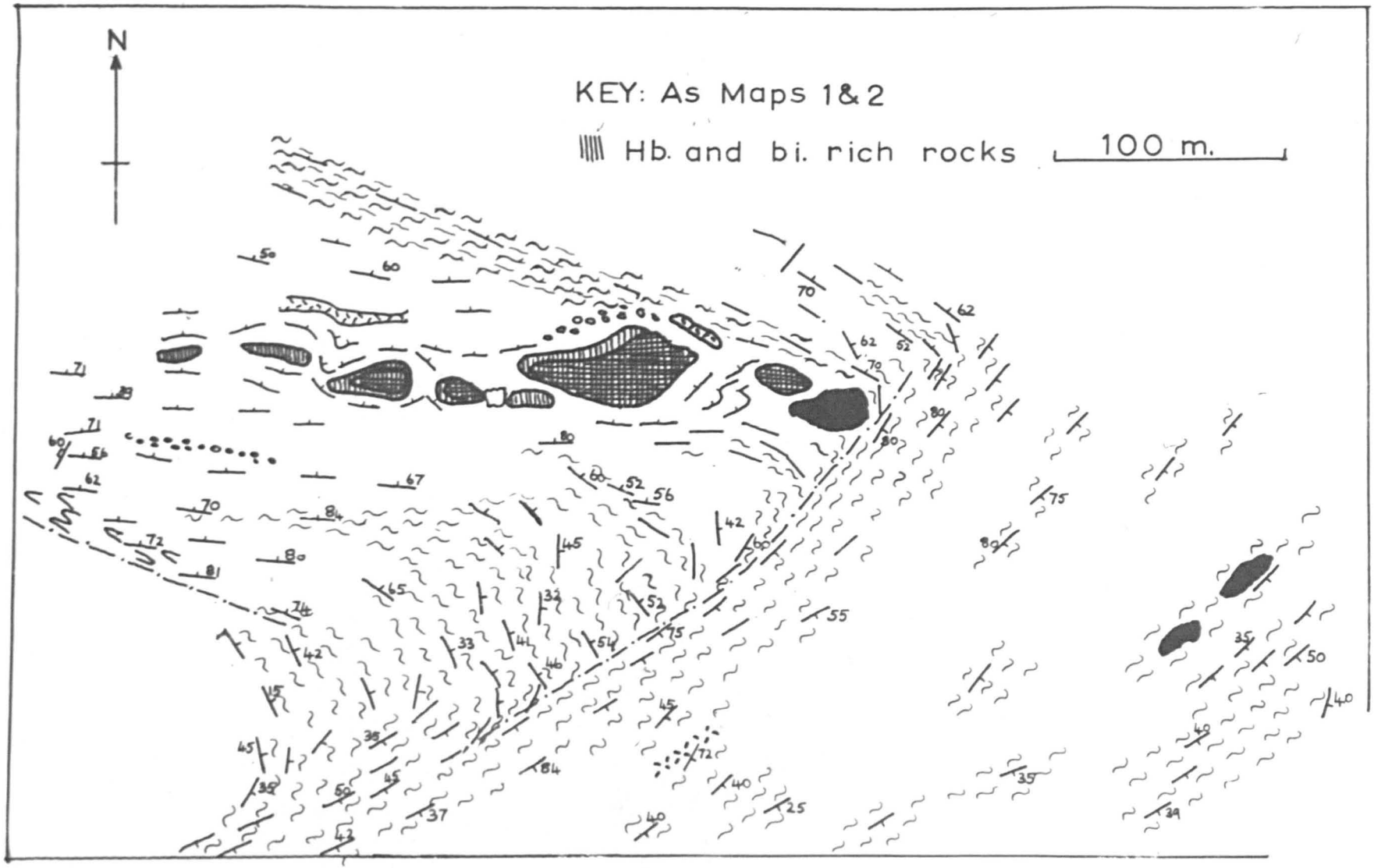


Fig.10: map of ultramafic bodies; Bousd, Coll.

succeeded at the margin by indistinct zones progressively richer in biotite and fibrous amphibole, until the outermost might be termed a biotitite. Small pods of bright green amphibole, about a metre across, are seen in the gneisses nearby. The other large outcrops are similar but are noteable in that they lie in narrow belts of possible metasediments. Green amphibole pods are found at wide intervals throughout the gneisses, possibly representing further ultrabasic bodies which have been dismembered and altered. At Rudha Hogh on Coll several pods of biotite are found in the nebulitic gneisses.

Agmatite belts frequently contain enclaves of hornblendite and pods of fibrous amphibole which represent the ultimate products of the complex process of agmatisation.

The final products of boudinage, deformation and further boudinage of metabasite bands in nebulitic gneiss are isolated black hornblendite pods.

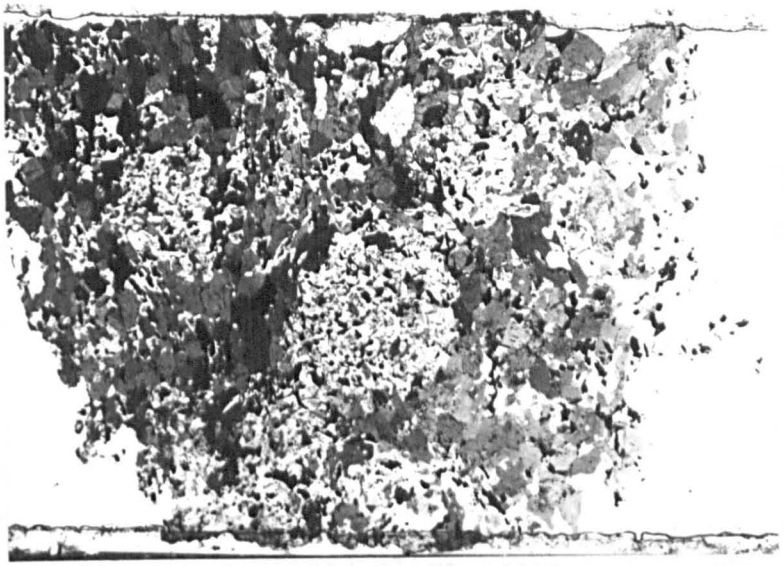
Primary ultramafic rocks.

Fresh undeformed ultramafic rocks from bands and lenses contain high grade assemblages. They are devoid of felsic minerals and range in grain size from 0.5 mm to 5 cm.

A prominent mineral present is strongly pleochroic pink to pale green orthopyroxene, occurring as rounded porphyroblasts (or phenocrysts) up to 5 cm across. They contain innumerable small exsolved blobs of ore, green spinel and pale green amphibole (Plate 24b), none showing epitaxic relationships with their host. They sometimes include corroded olivines. The margins of the orthopyroxenes are indented in contact with pale green amphibole in the matrix (Plate 25a). It is difficult to discern whether the amphiboles replaces orthopyroxene or they are simply in disequilibrium. The exsolved spinel indicates that originally orthopyroxene must have been highly aluminous, whilst the pale, exsolved amphibole might represent clinopyroxene originally in solid solution to a marked extent with orthopyroxene.

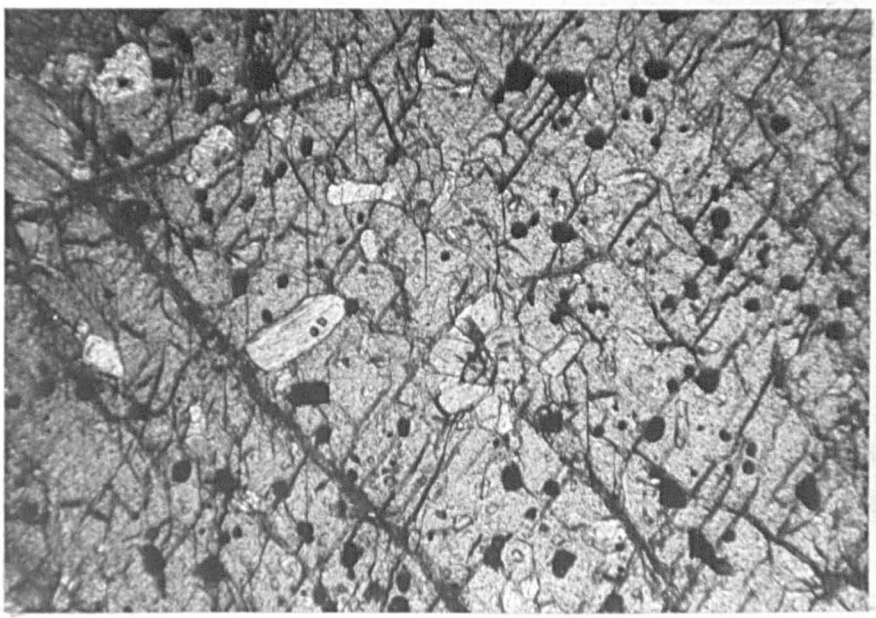
In the finer grained matrix are seen rounded and skeletal olivine grains, generally a little coarser than the surrounding polygonal network of pale green amphibole. In different samples they show various degrees of serpentinisation (Plate 25a) and may range from unaltered to small relics surrounded by colourless serpentine and ore.

The dominant mineral is pale green amphibole which forms the bulk of



a

Retrogressed garnets in amphibolite, pseudomorphed by fine-grained aggregate of plagioclase, hornblende and ore.
(Spec. 50676, p.p.l., x 3.5)



b

Orthopyroxene with exsolved ore, amphibole and spinel, in hornblende-peridotite.
(Spec. 5552B, x.p., x 60)

the matrix. It shows textures indicative of complete recrystallisation, i.e. polygonal aggregates with 120° triple points (Plate 25b). It is in disequilibrium with orthopyroxene and olivine. Several possibilities exist for its presence. It may have been a primary mineral subsequently recrystallised and changed in composition in response to its disequilibrium with the other phases or it replaced an earlier phase, probably clinopyroxene. The fact that no conclusive evidence for either is present precludes a choice.

Cummingtonite sometimes replaces orthopyroxene.

The hornblendite rim associated with the ultramafic body at Rubha Mor is a foliated equigranular rock from 1 to 5 mm in grain size, containing sparse plagioclase of An_{25} composition in lobate grains.

The dominant mineral is unsieved olive green hornblende in polygonal, equilibrated grains, which define the rock's fabric. Its associate is orthopyroxene in stubby rounded grains to which it does not show a reaction relationship. Curved contacts between the two minerals are usual (Plate 26a). Orthopyroxene is partly altered to biotite in places, but otherwise very fresh.

The outermost parts of the rim are progressively poorer in orthopyroxene and richer in felted actinolite and biotite, the outermost rim being composed of the latter two minerals (Plate 26b).

Ultramafic rocks associated with agmatites generally consist of polygonal aggregates of green hornblende with varying amounts of altered felspar. Fibrous amphibole layers are composed of tremolite-actinolite.

Discussion.

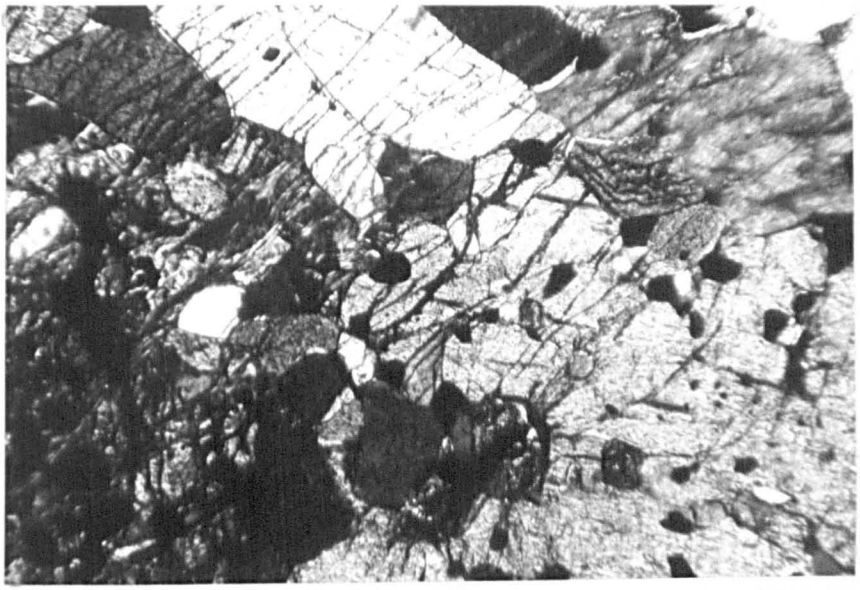
Ultramafic rocks of similar mineralogy are widespread in the Lewisian of the Mainland and the Outer Hebrides (Bowes et al. 1963, Davidson 1946).

The similarities are:

1) They are peridotites in a mineralogical sense.
2) They have a similar structural setting.
3) Both sets were present before a granulite facies metamorphism.

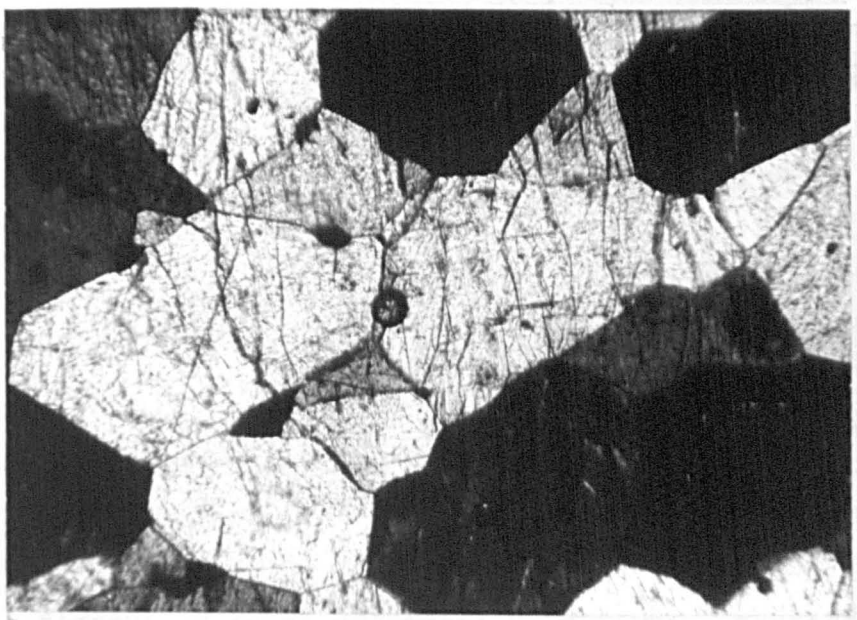
The essential differences are:

1) The fresh ultramafic rocks of Coll and Tiree are unbanded nor do they have a garnetiferous basic margin, thus they do not show evidence



a

Disequilibrium relations between orthopyroxene, pale hornblende and serpentinised olivine in hornblende-peridotite.
(Spec. 5552B, p.p.l., x 60)



b

Polygonal, recrystallised matrix of hornblende in hornblende-peridotite.
(Spec. 5552B, x.p., x 60)

for an origin as primarily banded intrusions (Bowes et al. 1963) or as tectonically and metamorphically banded rocks (Tarney, 1964; Sheraton, 1970) or as original dunites in a reaction relationship with country gneiss (O'Hara, 1961).

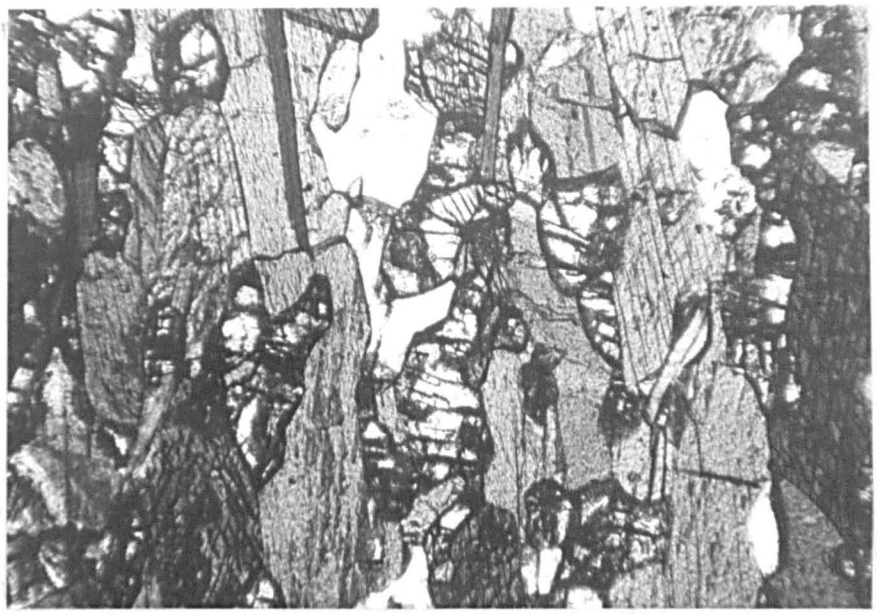
2) Spinel does not occur as a primary mineral in the fresh ultramafic rocks of Coll and Tiree, in contrast to the Mainland types (Bowes et al., 1963; Sheraton, 1970).

If the pale green amphiboles of the Coll and Tiree rocks are primary then they can be termed hornblende-peridotites. Conversely, if the amphiboles represent hydroxylated clinopyroxenes, then the rocks were originally two pyroxene-peridotites or lherzolites.

The lack of fabric in unaltered parts might represent igneous crystallisation or static recrystallisation under metamorphic conditions. The deformed rim of orthopyroxene-hornblendite is similar to the case with the Lizard complex (Green 1964) which was thought to be due to tectonic emplacement of a peridotite, in its case, causing localised granulite facies conditions. The assemblage hornblende-orthopyroxene-plagioclase might have formed under hydrous granulite facies conditions, but because of the ultrabasic composition orthopyroxene might not have been able to react with other constituents and remained stable in the amphibolite facies.

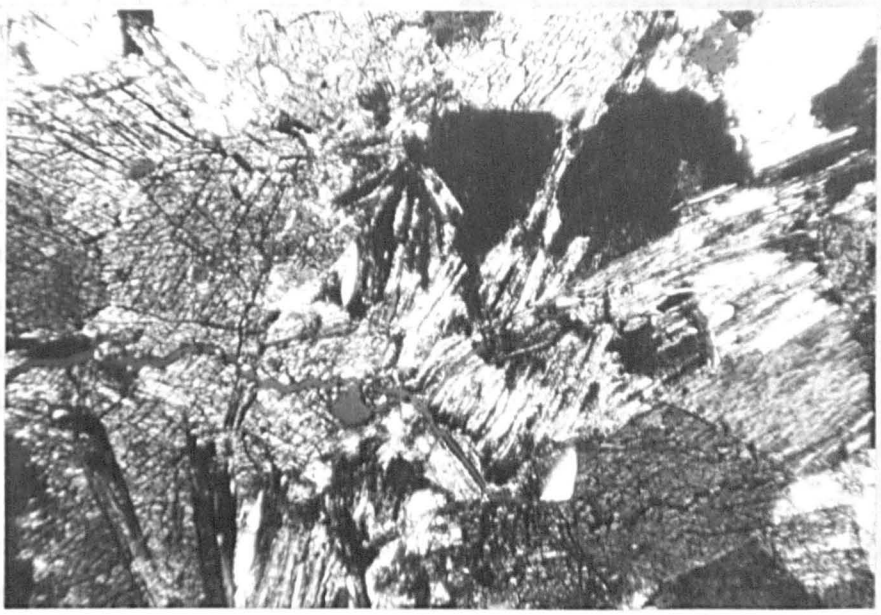
The outwardly succeeding biotite-rich portions perhaps indicate hydration with preferential acceptance of potassium during amphibolite facies metamorphism.

In the least altered parts, the stability of orthopyroxene with exsolved spinel, ore and possibly calciferous amphibole, suggests that earlier aluminous and calcic orthopyroxenes changed composition by exsolution in response to a later metamorphic event. Such highly aluminous and calcic orthopyroxenes are generally thought to be stable only under physical conditions at great depths, alumina representing part of that released by the breakdown of felspars in response to pressure (MacGregor 1967). O'Hara (1967b) studied the stability of various assemblages in artificial ultrabasic mixtures, similar to those naturally occurring. Specimen 78NW3B from Loch a' Phuil, Tiree has a norm (see Appendix 3 p. 209) comparing favourably with one of O'Hara's mixtures $MgSiO_4$ 40 per cent, $MgSiO_3$ 28.5 per cent, $CaMgSi_2O_6$ 28.5 per cent and Al_2O_3 3 per cent. For this mixture his experiments showed that plagioclase-lherzolite was converted to spinel-lherzolite at relatively low temperatures in the range 1000-2000°C with



a

Orthopyroxene with coexisting hornblende and plagioclase, rim to hornblende-peridotite at Bousd.
(Spec. 5552C, p.p.l., x 60)



b

Felted actinolite-biotite rock in outermost rim around hornblende-peridotite body at Bousd.
(Spec. 5552E, x.p., x 60)

increasing pressure in the range 0 to 50 kbar, and to lherzolite (pyrolite) at high temperatures (O'Hara 1967b Fig. 12.3p.394) when all alumina is contained in pyroxenes. Spinel-lherzolite and pyrolite change to garnet-lherzolite with further increases in pressure. O'Hara strongly stresses that the bounding pressure-temperature curves are critically dependant on composition. However, if the hornblende-peridotites of Coll and Tiree are partly hydrated pyrolite, then their primary crystallisation as ortho-pyroxene-clinopyroxene-olivine would undoubtedly have been governed by physical conditions only found in the mantle. This would only be verified by high pressure-temperature experiments.

The alternative case, that they primarily crystallised as hornblende-peridotites, with all alumina and calcium held in orthopyroxene and amphibole then they could still have formed in a regime of extreme physical conditions. O'Hara (1967a) indicates that water in ultrabasic mixtures at relatively low to moderate pressures and moderate temperatures (see O'Hara 1967a Fig. 1.4 p. 14) can only appear as vapour. At lower temperatures, however water may combine to form amphibole. The assemblage represented by the ultra-mafics of Coll and Tiree, aluminous orthopyroxene-amphibole-olivine, is apparently equivalent to O'Hara's hornblende-peridotite facies, and could therefore represent hydrated pyrolite. The bounding conditions for uni-variant equilibrium of such material in the system $CaO-MgO-Al_2O_3-SiO_2-H_2O$ (O'Hara 1967a Fig. 1.4 p. 14) are given as from 10 to 30 kbar and 800-950°C.

The marginal development of orthopyroxene-bearing hornblendic rocks with strong fabric is similar to the case with the Lizard peridotite (Green 1964) which has been attributed to tectonic emplacement of a peridotite, in its case causing localised granulite facies conditions.

The bearing of geochemical data on the origin of the ultramafic rocks is considered in Chapter 4.

Points of interest shown by primary ultramafic bodies are:

a) They were emplaced tectonically before granulite facies. Two are in thin metasediment bands.

b) They may retain relics of their original mineralogy. If so, they might represent material from great depths.

c) Their alteration in granulite and amphibolite facies metamorphism resulted in progressive hydration and enrichment in alkalies.

Granitic rocks.

Concordant granitic veins.

Pale, irregular granitic bands and veins are always to be found to some extent within granofelses, metabasites and garnet-biotite-gneisses. They are parallel to the margins of their host rocks, and range in size and amount from mere stringers less than 5 cm wide to 10 m thick bands exceeding the host rock in volume (Plate 27a). Their structural history is the same as that of concordant veins in quartzo-felspathic gneisses. "Feeder" veins are not seen, the granites are discontinuous, and they may have a genetic connection with their hosts.

Most granites are massive and devoid of foliaceous minerals. However, a planar element can be seen especially when the veins are thick or contained in foliated host rocks. A 5 m thick pink foliated granite with marked colour banding can be followed for about 2 km between Hogh Bay and Port na Luing (Map 1), a similar body being present from Eilean Bhoramuill to Feall Bay. Foliated pink granites are found with diopside-granofels south of Totronald next to a diopside-gneiss band in the main granofels. In mylonitised zones, the granite veins are still recognisable.

Whatever their host rock, there is no petrographic distinction between the granite veins, and they show a range in grain size from 0.5 mm to 2 cm, averaging 5 mm. They are typified by a granoblastic texture, foliation, where present, being defined by sparse mica and elongate aggregates of ferromagnesian minerals. Ferromagnesian minerals never exceed 1 per cent of the total volume.

Quartz is co-equal in amount with felspars, and typically in interlobate to amoeboid, strongly polygonised grains up to 2 cm across. In orthopyroxene-bearing varieties it has a polygonal habit. In lower grade rocks recrystallised polygonal aggregates lie along grain interfaces. In foliated veins, elongate aggregates of sutured quartz define a rough banding.

Alkali felspar is variable in amount, and from examination of stained material is in minor quantities in high grade rocks. It shows a marked increase in deformed varieties becoming the dominant felspar. In granulite facies veins, alkali felspar is largely untwinned and perthitic, showing hair-like and patchy exsolved sodic felspar. Such hair-perthites are typical of charnockites (Pichamuthu 1959). In the majority of veins

from amphibolite facies hosts it occurs as perthitic microcline, sometimes with exsolved quartz blobs. Development of microcline twinning was initiated in zones near cracks and boundary tumescences in untwinned alkali felspar, probably in response to strain (Spry 1969). Alkali felspar is present in antiperthites, and some of the larger microcline grains are undoubtedly due to recrystallisation of the exsolved material. Sen(1959) reports that natural plagioclases may contain up to 11 per cent dissolved alkali felspar in response to high grade metamorphic conditions, but not even complete exsolution could account for microcline being dominant over plagioclase in the most potassic veins. The consistently low alkali felspar content of granulite facies veins rules out that the variation was a primary feature. An attractive hypothesis is that plagioclase felspar has been replaced by microcline as a result of potash influx with decreasing grade and increasing deformation.

Petrographic evidence for this lies in several facts:

1) In the most potassic types, microcline is in polygonal aggregates containing skeletal plagioclase grains.

2) Plagioclase in contact with microcline has an An rich mantle in contrast to the case in migmatitic gneisses.

3) Some plagioclase-microcline boundaries have a film of quartz.

4) Microcline embays plagioclase and myrmekite is commonly associated, becoming a prominent feature in microcline-rich rocks.

Plagioclase is usually interlobate and rarely recrystallised, varying from An_{15} to An_{30}. It is commonly antiperthitic in high grade types, but less so in low grade potassic varieties. As suggested above, plagioclase has been replaced by microcline probably by grain boundary diffusion rather than nucleation within the plagioclase lattice. In low grade granites plagioclase is clouded with sericite possibly as a result of reaction between dissolved potash felspar and plagioclase at the onset of hydration. Some muscovite flakes may be due to metasomatism or recrystallisation of sericite.

Equilibrated ortho- and clinopyroxene are only present in the highest grade environments. They seem to have been replaced by garnet in some fashion.

Sparse porphyroblasts of garnet, up to 2 cm across are commonly present having a euhedral form, but sometimes of an atoll type. They are inclusion-free. Small flakes of biotite mantle them, perhaps in a re-

placive mode. In some deformed veins, garnet has been granulated and small angular grains are in streaks along the foliation. Yellow-brown biotite is sparse, and may be aligned with a planar fabric. It is altered to chlorite and prehnite.

Accessories include apatite, ore and zircon.

Discussion.

The fact that some of the concordant granitic veins contain orthopyroxenes and they show a structural history as complete as their host rocks, suggests that they were formed prior to the granulite facies metamorphism at some stage during it. Their process of formation could have been a result of several possible processes, or a combination of processes:

1) As intrusive granite.
2) As metasomatic secretions resulting from magmatic activity.
3) By a process of metamorphic differentiation.
4) By partial melting or partial solution by hydrous fluids of their host rocks.

Cases 1 and 2 are not supported by the presence of large granite bodies of a similar age.

Case 3 is precluded by the lack of foliation in most of the granites, and their large but variable volume and irregularity.

Even in the absence of melanocratic rims at host-granite boundaries, Case 4 is the most consistent with observed features. Further discussion is left until Chapter 4.

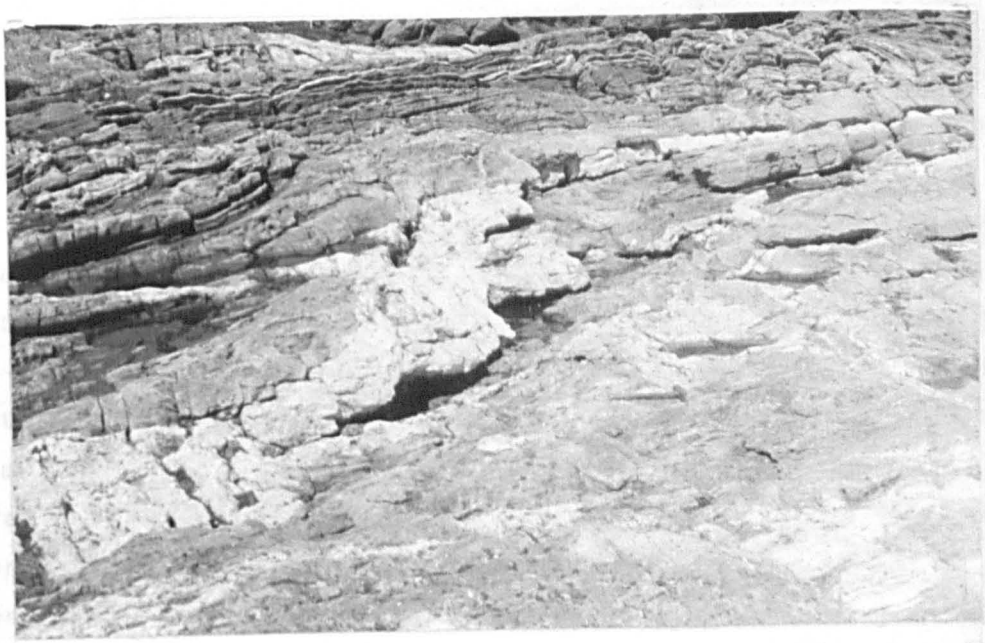
The metamorphic history of these rocks is similar to other members of the Lewisian complex of Coll and Tiree. A noteable feature is the apparent increase in alkali felspar content with decreasing grade and increasing deformation. The hypothesis that this is due to potash metasomatism during the amphibolite facies is returned to in Chapter 4.

Discordant pegmatites and granites.

Very distinctive white quartzose pegmatites are seen cutting the migmatitic hornblende-biotite-gneisses of Coll, which are up to 25 m thick. Undeformed examples show a zoned nature, a sometimes rose-tinted pure quartz core being roughly margined by a coarse (2cm) quartz-felspar rock. They are thus zoned pegmatites. Several examples transgress the foliation

a

Concordant granite veins in granofels at Sloc na Stuir.



b

Remobilised white granite with related concordant granitic veins in amphibolites and granofelses at Bagh na Coille.

and F_2 axial surfaces and they have been folded ptygmatically during D_3 (Plate 4d). The most are concordant showing varying degrees of deformation. Prominent examples outcrop on Meall Eatharna and between Acha and Clabhach.

White granitic pegmatites, similar in appearance to concordant veins in various rock types, but transgressive to the foliation, are seen at Balephetrish on Tiree and at Sorisdale on Coll (Plate 27b). At Sorisdale they partly follow F_5 steep limbs and are cut by pink biotite-granites. At Balephetrish a 150 m thick body cutting metabasite is seen to be made up of white granite without mafic minerals grading into pink biotite-granite.

Large dykes and pipe-like bodies of pink biotite-granite or pegmatite up to 10 m across are affected only by F_6 shears. They can be extremely coarse (up to 10 cm) and made up of a network of intersecting thin biotite flakes separated by felspars and quartz (Plate 28a). They outcrop as boat-shaped roches mouttonées. Where affected by F_6 shearing, and mylonitisation they are bright red and chloritic.

Quartzose pegmatites.

These intrusive rocks show variable textural products of deformation. Grain size varies with the deformed state from 5 mm in the undeformed to 0.05 mm in the most deformed.

Quartz is dominant (up to 90 per cent), and shows strong undulatory extinction and sutured sub-amoeboid grain boundaries. The subgrains of deformed origin define a planar fabric, and in some cases have recrystallised as elongate aggregates. In other cases, quartz, has been granulated to a mozaic of sutured interlocking grains less than 0.1 mm across. Such fine grained quartz has a fabric defined by recrystallised plates and by narrow belts of even finer granulation, and which is deflected around felspar augen.

Plagioclase felspar is present as lenticular augen and has an oligoclase composition. Lamellar twins are bent and may have a spiral appearance. Plagioclase is usually clouded and contains abundant sericite.

Untwinned alkali felspar is in minor quantities, and like plagioclase, is in the form of augen, but often showing undulose extinction. Muscovite flakes are often seen within the alkali felspar and closely associated with it, lying in the foliation of the granulose quartz matrix.

Accessory minerals are generally absent.

These pegmatites are therefore intrusive, dominantly quartz with mainly sodic plagioclase and only a little alkali felspar.

From compositional considerations, Bowes (1967) held that certain autochthonous quartz-plagioclase veins in the Lewisian were products of supercritical hydrous solution in the amphibolite facies. The age of the Coll material is between D_2 and D_3 when conditions were perhaps changing from granulite to amphibolite facies, and hydrous fluids may have begun to permeate the complex.

Discordant white granites.

These rocks are mineralogically very similar to the concordant white granitic veins described above, except that they do not contain garnet. Their alkali felspar content is similar to that of those veins which are garnet free and deformed. They are, however, generally coarser with a grain size between 2 and 5 mm, and show straight grain boundaries and 120° triple points in patches of polygonal quartz and felspar, indicative of partial, equilibrium crystallisation. Their similarities to the concordant granites and their presence in areas of granofels, metabasite and garnet - gneiss suggests they were mobilised material derived from the veins in these rocks after D_5.

Biotite-pegmatites and granites.

Apart from grain size, both rock types are petrographically and structurally identical, being post-D_5 and pre-D_6. Grain size seems to increase with the size of the bodies, large outcrops having crystals over 10 cm in extent, and small pipes and veinlets being of aplo-granitic texture. They consist of mainly quartz, alkali and plagioclase felspars in roughly equal amounts with lesser quantities of biotite.

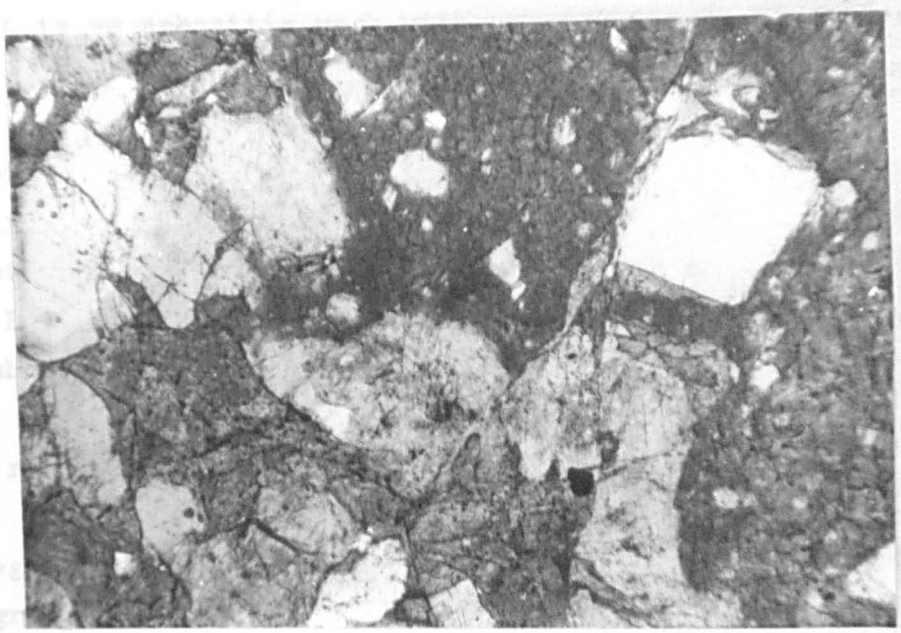
The macroscopic polygonal network texture of biotites (Plate 28a) is suggestive of crystallisation from a fluid.

Quartz is usually in strained, interlobate grains, but may show equilibrium textures.

Alkali felspar is present as untwinned vein and patch perthite with

a

Biotite network in biotite-pegmatite, Port na Luing, Coll.



b

Flinty crush rock with pumpellyite pseudomorphs after biotite.
(Spec. 4024, p.p.l., x 60)

exsolved quartz in some samples, showing only incipient microcline twinning, when its texture is interlobate. Other samples contain microcline in equilbrated polygonal aggregates with quartz.

Plagioclase of average An_{30} composition is in polygonal grains in association with untwinned alkali felspar. With microcline it has largely been converted to myrmekite. Most plagioclase grains are clouded and sericite may be prominent.

Biotite is the only prominent mafic mineral and has a random network distribution. It is frequently chloritised and in shear belts replaced by chlorite completely.

Accessories are apatite, ore and zircon. In one specimen (5573B) brown allanite is prominent.

The biotite-granites and pegmatites were originally magmatic rocks, and the gradation into white granite at Balephetrish on Tiree may indicate a genetic connection with earlier concordant granites.

Mylonitic rocks.

Black flinty crush rock occurs as irregular anastomosing veins in the gneisses. It is an aphanitic rock with porphyroclasts of gneiss, quartz and felspar. In thick veins it often has a brecciated appearance (Plate 11d).

Green epidotic crush rock, later than flinty crush rock, is aphanitic and breccia-like, It may occupy irregular veins or be aligned along wrench faults, thrust planes and joint faces. An arcuate dislocation between Bhasapol and Eilean Ghreasamuil on Tiree is marked by a 100 m zone of epidotic crush rock and breccia.

Flinty crush rock.

The irregular, sharp sided veins of this rock consist of almost isotropic fine grained material devoid of fabric. Set in it are fragments of quartz, plagioclase and microcline, all showing kinks and microfractures (Plate 28b). Deformation subgrains give them an undulose extinction.

The fine grained material is largely submicroscopic and yellow-brown. In strain shadows near porphyroclasts are aggregates of yellow-brown, lightly pleochroic, foliaceous pumpellyite which has typical undulatory

oblique extinction and fairly high relief (Plate 28b).

Mineral grains adjacent to the flinty veins show strong grain boundary granulation and biotite is pseudomorphed by pumpellyite and prehnite.

Flinty crush rocks are probably devitrified glasses as evidenced by their intrusive habit. Their angularity and content of strongly fractured relics indicates that they were products of a high rate of strain. The veins never show planar fabrics, are not associated with faults, and the host rock is rarely dislocated more than a few centimetres. They might have formed by some "explosive" mechanism, filling cracks in the rock opened by the sudden redistribution of stresses. It has been suggested that there may be some relationship between the vitreous base of flinty crush rocks and glass bearing impactites (Spry 1969), but those of Coll, being devitrified, cannot show the presence of thetomorphic silica, nor do they contain coesite or stichovite. They may well have formed as a result of some very rapid release of load pressure late in the history of the gneisses.

Epidotic crush rock.

Later than the flinty crush rocks, epidotic fine grained mylonitic material shows a few features in common with them. They are sometimes of sub-microscopic grain size and contain fractured and bent porphyroclasts. However, they have largely recrystallised to almost pure granular epidote. Colour banding is indicative of translation, which is to be expected from their occurrence along D_7 dislocations of every magnitude. Nearby gneisses are pink and granulated in bands parallel to small veins. As the epidote crush veins are monomineralic some process of ionic transfer must have contributed to their formation. Whether they were originally vitreous or merely finely granulated is obscure, but their association with some hydrous fluid would seem most probable.

Discussion.

The relationship of metasediments on Coll and Tiree to the now almost thoroughly downgraded quartzo-felspathic gneisses is obscurred by similar effects on the metasediments which exist in outcrops of three different scales. Metasedimentary series with similar lithological content and

history are widespread throughout the Lewisian of Scotland. Perhaps the most comparable series is that described by Coward et al. (1969) from the Outer Hebrides, where metasediments outcrop in a similar variation of scale. They have shown that the metasediments were present before Scourian granulite facies metamorphism and were retrogressed and partially migmatised during subsequent events. They suggest that their close association with metabasic rocks has in some way protected them from complete migmatisation during their history. The relationships between metasediments and gneisses are accounted for by two contrasting hypotheses. First, that the sediments and gneisses represent a congruent sedimentary pile, sediments only remaining distinct because of their composition or situation. Conversely they suggest that the sediments are a cover to a basement of earlier gneisses, the sedimentary belts representing tightly appressed synforms modified by multiple orogenesis. The metasediments of Coll and Tiree are identical in most respects to the Outer Isles series and similar hypotheses may be offered for their origin. The continuous metasediment belts probably represent original wide belts of distinctive sediment, whilst thinner bands and pods show the effects of break up of thin intercalations within the parental material of the migmatitic gneisses.

Prior to granulite facies metamorphism, the sediments and gneisses were intruded by basic and intermediate rocks, and small ultrabasic masses were emplaced tectonically. The undeformed ultrabasic rocks show mineralogical similarities to hypothetical rocks thought by some authors to be present at great depths. Granofels and metabasite bands are frequently associated with metasediment belts (Map 1) and three hypotheses can be suggested for this observed relationship. Either metasediments remain recognisable only where they have been protected from the full effects of metamorphism by granofels and metabasite bands, or intrusion was to some extent controlled by the presence of distinct inhomogeneities in the pre-granulite facies terrane provided by metasediment bands. The final possibility is that some of the granofels and metabasites represent volcanogenic rocks associated with the sediments. The Balephetrish Hill metabasite (Map 3) contains numerous enclaves of marble, and it would appear to have been intrusive into the Balephetrish - Hynish Bay belt. In the author's opinion, the most likely mode of origin for the concordant metabasites would be as intrusive sills, or as discordant bodies rotated into parallism by multiple orogenesis.

Some inferences can be drawn from the early features shown by hypersthene-gneisses of Western Tiree. The presence of segregation banding suggests that they had been through a period of migmatisation prior to their being intruded by intermediate and basic materials. Migmatites are generally accepted to be the product of an amphibolite facies hydrous environment when various sorts of fluids encourage ionic migration and a general redistribution of elements in rocks in response to orogenic forces. Sheraton (1970) has produced strong geochemical evidence that the two pyroxene-gneisses of Drumbeg suffered amphibolite facies migmatisation prior to the Scourian granulite facies. It would be reasonable to suggest that the hypersthene-gneisses of Tiree and their reworked products suffered a similar metamorphism at some stage before granulite facies. The presence of sillimanite as helycitic trails in deformed garnet-biotite-gneiss may be more direct evidence for the postulated event.

Several hypotheses can be erected concerning the physico-chemical conditions pertaining during the granulite facies metamorphism. The presence of kyanite in preference to sillimanite in pelitic gneisses at granulite facies suggests that pressure was above that necessary for the inversion of sillimanite to kyanite. Green and Ringwood (1967) have suggested that kyanite-bearing pelites in granulite facies terranes may indicate an intermediate to high pressure subdivision of the granulite facies, sillimanite being the alumino-silicate typical of intermediate pressures, and cordierite of low pressure granulite facies. The presence of brown hornblende, and possibly biotite, as phases in equilibrium with other granulite facies minerals, might indicate incomplete dehydration by the metamorphism, the fugacity of water varying, depending on individual bands' composition. Tarney (personal communication) has suggested that amphibolite facies hornblendes, changing composition with increasing grade, may become stable in the granulite facies, holding water in the host rocks.

Field, structural, textural and mineralogical evidence suggests that some concordant granitic veins in granofelses, metabasites and garnet-biotite-gneisses, were formed either before or during granulite facies metamorphism.

The possible reaction:

orthopyroxene + plagioclase + ore $\rightleftharpoons$ garnet + clinopyroxene + quartz,

is demonstrable in orthopyroxene-bearing granofelses and metabasites. It was initiated after the granulite facies crystallisation of assemblages

consisting of quartz, plagioclase, clinopyroxene, orthopyroxene and brown hornblende, and was anhydrous. In metabasites it has tended to produce an eclogite-like assemblage. Green and Ringwood (1967) have experimentally shown that in basic rocks a granulite facies assemblage, containing orthopyroxene and plagioclase, would change to an eclogite-like assemblage of garnet and clinopyroxene with decreasing temperature at constant pressure, albeit sluggishly. It is suggested here, that such a metamorphic event occurred in the rocks under study, being interrupted by amphibolite facies metamorphism before coming to completion. The pertinence of major element geochemistry to this garnet producing reaction and the associated physical conditions will be covered in Chapter 4.

The presence of hydrous phases, such as green hornblende and biotite, in many samples from both Coll and Tiree indicates that amphibolite facies metamorphism after the granulite facies event, retrogressed the earlier assemblages, almost completely, its effects increasing from west to east. As the state of deformation also increases from west to east, it seems probable that incoming hydrous fluids influenced deformation during the amphibolite facies.

Evidence from granofelses and metabasites indicates that, during the amphibolite facies metamorphism, conditions when almandine garnet, hornblende, diopside and biotite were stable gave way to lower grade conditions marked by the presence of hornblende and biotite only.

Throughout the history of the retrogressive metamorphism such phases as garnet, hornblende and biotite can be seen to have formed in a multiplicity of ways at different times in one particular rock type. In some samples, mineral assemblages of up to 7 or 8 phases are present, and one particular phase may be demonstrated to be of several generations. Saxena (1969) has stated that in many mineral reaction equations defining the transition from amphibolite facies to granulite facies (De Waard 1965, 1966) solid solution phases are present on both amphibolite and granulite facies "sides" of the equation. All phases may be present in equilibrium under particular physical conditions, definite relationships existing between the chemistry of multi-component phases. In fact he suggests that the presence of all reacting phases in a rock is necessary for it to be possible to tell whether a particular reaction has occurred, and that electron probe analyses are essential to ascertain whether equilibrium has been attained. A further possibility is that numerous reactions,

having taken place in the various rock types of Coll and Tiree, proceeded at such a slow rate that physical conditions changed before equilibrium could be reached. This might have resulted in the observed series of mineral species being preserved in a metastable state, showing complex textures of several generations. In the case of solid solution minerals, such as plagioclase, hornblende, biotite and garnet, it is possible that different generations of particular minerals are now chemically similar, chemical equilibrium having been maintained via an intergranular fluid or disperse phase. This being the case, it might be expected that co-existing solid solution minerals display regular chemical relationships to one another with respect to major elements. This topic is returned to in Chapter 4.

The culmination of the amphibolite facies, retrogressive metamorphism was the mobilisation of quartzo-felspathic materials resulting in the formation of nebulitic gneisses and agmatites. Textural evidence from concordant granitic veins in granofelses and metabasites suggests that they were progressively enriched in potassium during the amphibolite facies.

The lithologies and structures produced were transgressed by large dyke- and pipe-like biotite-pegmatites, which may have been produced by partial melting. Some are demonstrably derived from remobilised concordant granitic rocks.

Shearing after the intrusion of biotite-pegmatites produced a new foliation and a greenschist facies mineralogy in gneisses, granofelses and amphibolites. It has been shown that this further retrogression was associated with an orogenic phase.

The latest metamorphic activity was a double period of mylonitisation, strongly developed on the NE coast of Coll. The first mylonites were black flinty-crush rocks developed with associated brittle folds parallel to monoclinal steep belts. They are similar to pseudotachylites described by numerous authors from the vicinity of the Outer Isles Thrust (Fig. 1) and from the Lewisian of the Gairloch region described by Park (1961). They are cut by green epidotic-crush rocks associated with thrust planes and wrench faults. They are similar to mylonites described in the Lewisian by the Survey (1920) from Iona and Islay which were thought to be partly post-Torridonian in age.

TABLE 2 : Partial analyses of felspar/quartz fractions

	1	2	3	4	5	6	7
SiO_2	75.45	79.11	68.7	77.5	58.3	77.9	56.7
TiO_2	-	-	-	-	-	-	-
Al_2O_3	15.5	13.7	17.2	14.8	23.7	13.7	27.4
Fe_2O_3	-	-	-	-	-	-	-
MnO	-	-	-	-	-	-	-
MgO	-	-	-	-	-	-	-
CaO	3.27	2.55	1.85	3.82	6.95	3.38	10.43
Na_2O	3.7	3.7	3.2	4.2	6.8	4.1	5.3
K_2O	1.67	2.49	6.45	0.39	0.399	0.443	0.459
Trace elements in parts per million							
Rb	68	56	148	8	6	15	6
Sr	370	305	177	490	758	267	429
Y	1	2	3	1	1	1	1
Zr	119	50	148	36	64	71	60
Nb	1	1	1	1	1	1	1
Ba	357	437	1154	235	287	184	197
La	4	2	1	2	13	1	1
Ce	5	2	3	1	21	1	8
Pb	22	50	10	46	21	45	14
Th	4	6	2	1	1	1	1

1. Mean of 6 analysed fractions from garnet-biotite granofels.
2. Mean of 8 analysed fractions from concordant granites.
3. Felspar/quartz fraction of diopside-gneiss 7321I.
4. Felspar/quartz fraction of garnet-biotite-gneiss 4965A.
5. Mean of 2 analysed fractions from mafic granofelses.
6. Felspar/quartz fraction from garnet-amphibolite C1.
7. Felspar/quartz fraction from pyribolite 78SW5B.

TABLE 2 (cont.) : Partial analyses of biotites and garnets

	1	2	3	4	5	6	7
SiO_2	-	-	-	-	-	-	-
TiO_2	2.94	3.34	0.09	0.07	0.12	0.15	4.00
Al_2O_3	16.1	14.1	18.2	17.5	18.4	16.1	18.4
Fe_2O_3	25.40	27.75	32.99	32.98	33.01	31.51	30.69
MnO	0.06	0.09	1.19	1.27	1.57	1.21	1.57
MgO	14.4	14.7	6.1	6.1	6.3	4.4	3.0
CaO	3.82	2.26	3.65	2.27	3.38	6.91	8.8
Na_2O	0.4	0.1	-	-	-	-	-
K_2O	4.26	7.98	-	-	-	-	-
Trace elements in parts per million							
Cr	1500	675	714	497	488	356	712
Ni	1120	501	241	197	185	120	200
Rb	128	286	-	-	-	-	-
Sr	94	73	43	36	44	46	1
Y	10	40	58	413	224	105	83
Zr	66	107	28	29	26	15	32
Nb	13	22	15	10	11	8	80
Ba	797	1493	147	231	223	166	252
La	88	121	100	84	197	65	57
*Ce	28	300	1	1	1	1	61
Pb	25	55	6	14	19	2	1
Th	4	16	4	8	7	1	5

* Ce concentration lowered by effect of high Cr on background x-ray intensity for Ce.

1. Mean of 2 biotite analyses from garnet-biotite-gneisses.
2. Mean of 6 biotite analyses from garnet-biotite-granofelses.
3. Mean of 2 garnet analyses from garnet-biotite-gneisses.
4. Mean of 6 garnet analyses from garnet-biotite-granofelses.
5. Mean of 4 garnet analyses from concordant granite veins.
6. Analysis of garnet from mafic granofels 78NW1B.
7. Analysis of garnet from garnet-amphibolite C1.

TABLE 2 (cont.) : Partial analyses of garnet, pyroxenes and an allanite

	1	2	3	4	5
SiO_2	-	-	-	-	
TiO_2	0.07	0.12	.634	2.561	
Al_2O_3	15.0	4.1	2.8	3.9	
Fe_2O_3	30.72	23.00	11.20	6.53	
MnO	0.666	0.02	0.14	0.44	
MgO	6.0	41.16	16.00	12.83	
CaO	6.63	0.54	21.74	20.88	
Na_2O	-	-	-	0.75	
K_2O	-	-	-	-	
Trace elements in parts per million					
Cr	290	675	194	37	1
Ni	95	651	65	1	1
Rb	-	-	-	-	-
Sr	37	82	5	10	837,
Y	82	-	9	6	125
Zr	18	4	17	41	107
Nb	6	90	1	2	19
Ba	164	221	127	123	60153
La	84	1	3	6	32911
Ce	1	1	35	30	45576
Pb	1	1	22	31	1
Th	5	9	3	1	3029

1. Analysis of garnet from pyribolite 78SW5B.
2. Analysis of orthopyroxene from hornblende-peridotite 5552B.
3. Analysis of clinopyroxene from diopside-gneiss 7321I.
4. Analysis of clinopyroxene from pyribolite 78SW5B.
5. Analysis of allanite from biotite-pegmatite 5573B.

TABLE 2 (cont.) : Partial analyses of hornblendes

	1	2	3	4	5	6
*SiO_2	49.6	50.8	42.5	40.5	44.0	48.8
TiO_2	0.28	0.27	2.08	1.70	1.54	.367
*Al_2O_3	4.0	7.0	13.0	10.5	12.6	5.8
Fe_2O_3	22.81	10.41	16.37	12.47	19.60	7.94
MnO	0.48	0.39	0.23	0.14	0.30	0.13
MgO	18.0	18.0	10.75	9.4	9.0	25.4
CaO	0.40	13.17	10.61	8.24	12.00	12.47
Na_2O	0.51	1.1	1.7	1.0	1.7	0.8
K_2O	0.112	0.535	1.675	1.315	0.556	0.319
Trace elements in parts per million						
Cr	417	57	184	347	409	1893
Ni	28	1	177	163	168	1262
Rb	2	11	15	4.5	13	3.1
Sr	1	2	35	1	22	45
Y	7	22	7	1	24	15
Zr	22	15	68	1	37	48
Nb	1	1	4	7	1	6
Ba	101	167	439	362	215	88
La	3	21	27	30	2	5
Ce	1	59	62	65	4	1
Pb	1	69	1	1	36	19
Th	1	10	8	21	1	4

* SiO_2 and Al_2O_3 analyses for hornblendes are regarded as dubious and are imprecise.

1. Analysed anthophyllite/gedrite from anthophyllite-gneiss 70815E.
2. Analysed actinolitic hornblende from diopside-gneiss 7321I.
3. Analysed brown hornblende from mafic granofels 78NW1B.
4. Analysed brown hornblende from pyribolite 78SW5B.
5. Analysed green hornblende from garnet-amphibolite C1.
6. Analysed pale green hornblende from garnet-peridotite 5552B.

CHAPTER 4

GEOCHEMISTRY

210 whole rock specimens were analysed using a Phillips PW1212 X-ray spectrometer. A short account of sample preparation and analytical procedure are given in Appendix 1, where the precision and accuracy of the method are discussed. Mineral assemblages for all the analysed rocks are given in Appendix 2. Listed individual analyses are contained in Appendix 3. Figs. 11 and 12 show the sample localities on Coll and Tiree.

The object of the geochemical study was fourfold:

1. To collect data regarding the origins of various rock types.

2. To estimate the effect of major element chemistry on mineralogy of various rock types at different grades of metamorphism, and the effect of mineralogy on trace element abundances.

3. To investigate possible chemical changes during the metamorphism of various rock types, and to establish which elements are fractionated by the processes of regional metamorphism.

4. To compare the chemistry of the dominant gneissose rocks of Coll and Tiree with average analyses from elsewhere in the Lewisian for correlation purposes, and with averages of Precambrian basement rocks from other parts of the world.

The second objective was studied by comparing average analyses of small groups of mineralogically alike samples from the larger groups of particular rock types, and by interpreting inter-element correlation coefficient matrices for the small groups. Several different common minerals were also analysed by X-ray spectrometry to this end.

In evaluating metamorphic fractionation processes in the complex, correlation coefficient matrices for certain elements and critical element ratios were generated for each compositionally limited major group of samples.

The correlation matrices and significance levels were generated by a program written by A.C. Skinner (1970), using the University of Birmingham KDF 9 computer.

For purposes of description and discussion, the analyses are presented in four main sets:

Gneisses and metasediments.

Possible intrusive igneous rocks; including metabasites, granofelses, and ultramafic rocks.

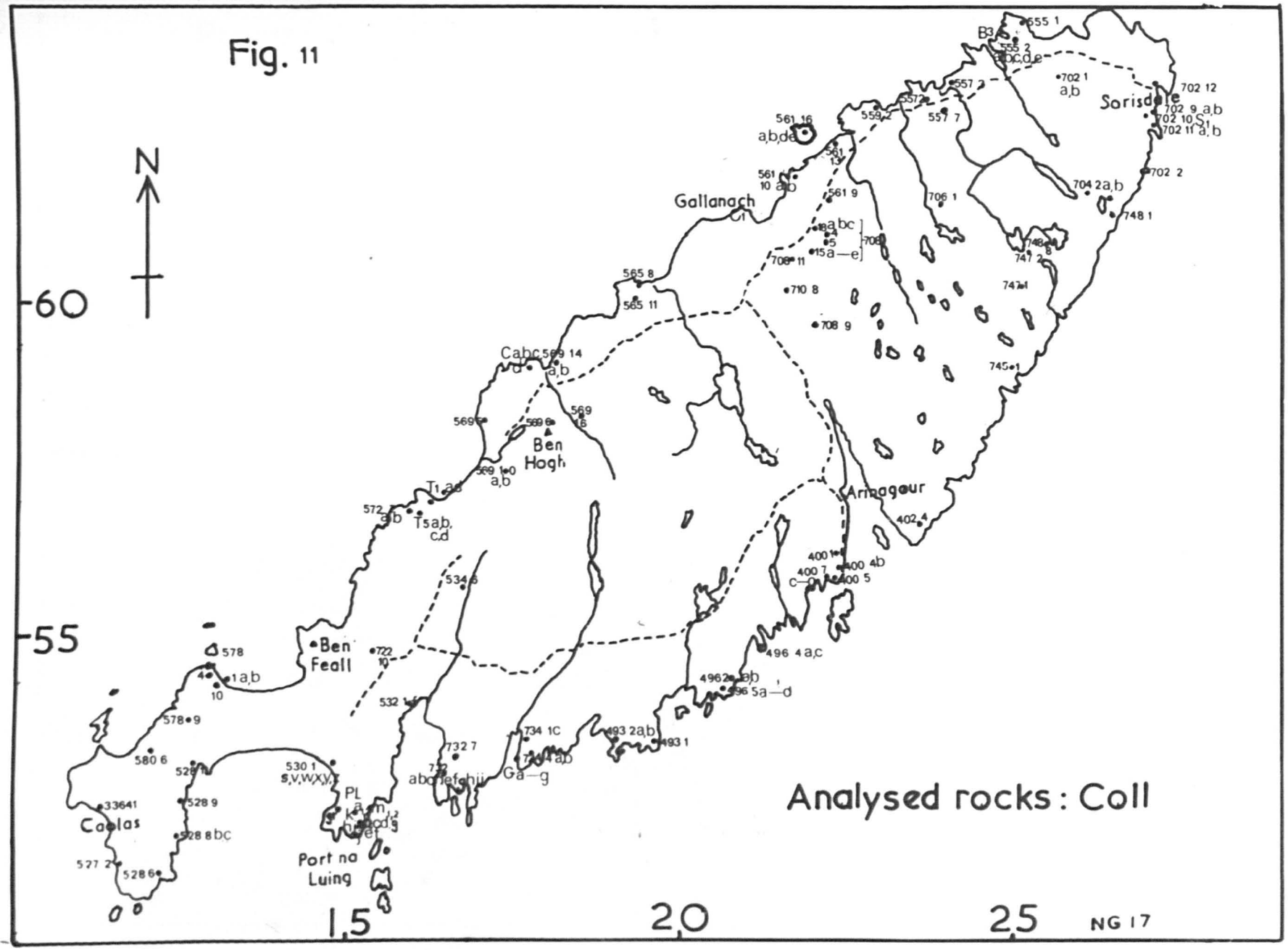


Fig. 11
N
60
55
Gallanach
Ben Hogh
Ben Feall
Arinagour
Soriscale
Port na Luing
Caolas
Analysed rocks : Coll
1.5
2.0
2.5
NG 17

Granitic rocks.

Mylonitic rocks.

Each group is described geochemically, with emphasis on trace element abundances, and the members are compared with one another and similar rock types from elsewhere. Possible origins for each distinctive lithology are discussed. Metamorphic fractionation trends for the whole complex are described, and their significance in the evolution of basement complexes is discussed.

In view of the fact that none of the rocks from Coll and Tiree contain original minerals, except perhaps the fresh ultramafic rocks, the trace element and, to some extent, the major element distributions are the products of metamorphism. Any original relationship must have been masked by a certain amount of element redistribution, and the trace elements at least will reflect the limits of their possible substitution for major elements in metamorphic minerals present.

Gneisses and metasediments.

Average compositions for the five distinguishable groups of gneiss are given in Table 3. Table 5 contains the means of several analyses of garnet-biotite-gneisses and anthophyllite-gneisses. All rock types under this heading are plotted together on several bivariate graphs and triangular diagrams, on which they have been represented by different symbols. Analyses of rocks of similar composition from different areas are given in Tables 4 and 5 for purposes of comparison.

Quartzo-felspathic-gneisses.

The composition fields of all the recognisable groups, except diopside-gneisses, have similar extents and overlap on An-Ab-Or, Fe-Mg-Alk and ACF diagrams (Figs. 13a, 14a and 15a). Similarly with plots of Niggli k against mg (Fig. 16a).

On an ACF diagram (Fig. 15a) the hypersthene-gneisses do not all fall within the plagioclase-clinopyroxene-orthopyroxene field of the granulite facies, showing a spread towards the A apex, some falling in the garnet-orthopyroxene-plagioclase field. As all contain pyroxenes, frequently both monoclinic and orthorhombic, and none carry garnet, this perhaps

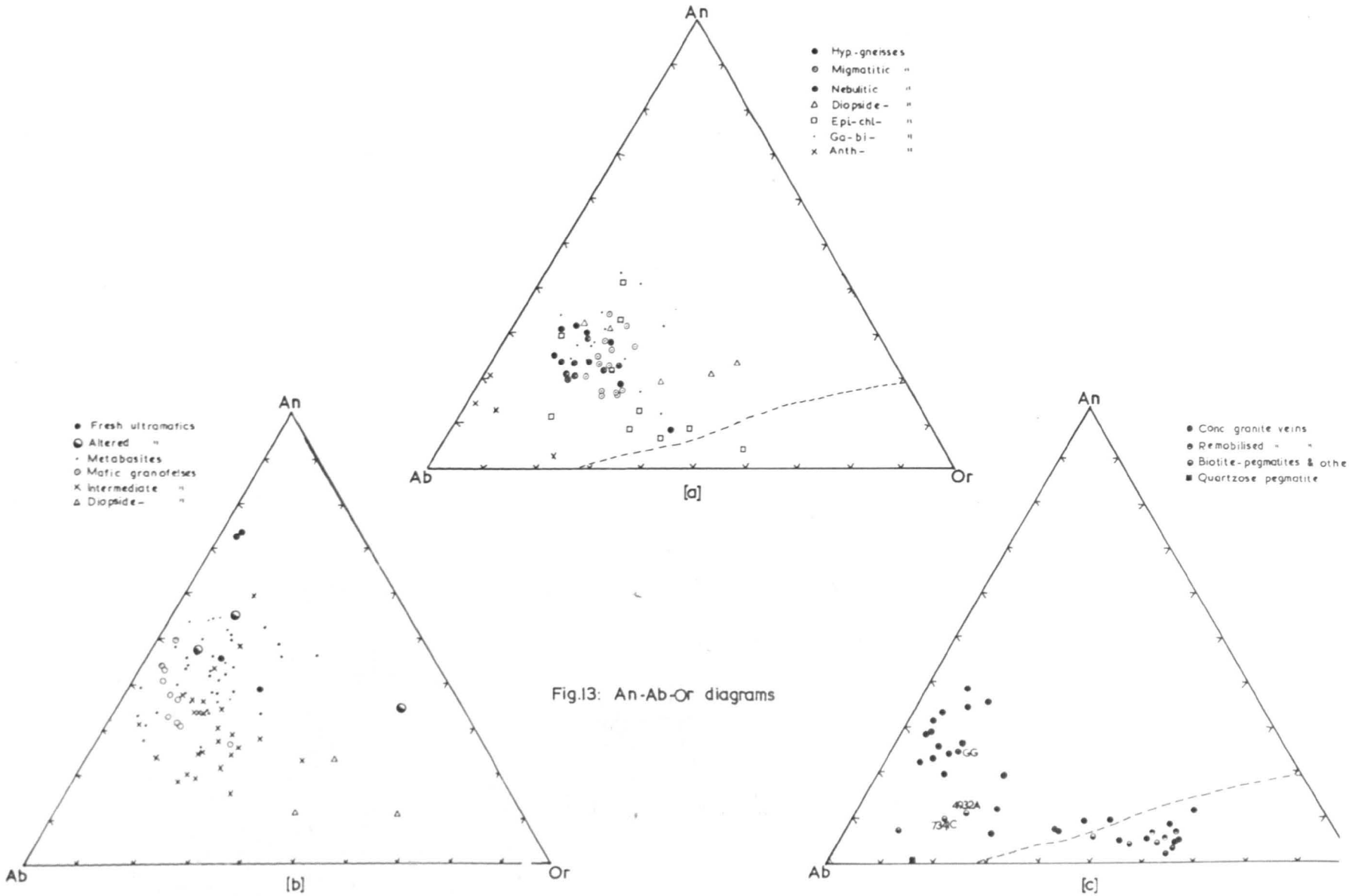


Fig.13: An-Ab-Or diagrams

illustrates the hypersensitivity of such a diagram to small differences in composition, especially in acid types. The observed spread might be due to correction for ore minerals not having been carried out. The diopside gneisses plot as they should, in the plagioclase-hornblende-clinopyroxene field of the amphibolite facies (Turner, 1968). Their relative position on the ACF plot shows their high CaO and MgO compared with the other gneisses.

The Fe-Mg-Alk plot (Fig. 14a) shows a strong linear trend for all the gneisses except the diopside-gneisses, which plot towards the Mg apex. In the amphibolite facies gneisses, and the greenschist facies epidote-chlorite-gneisses the prominently constant Fe/Mg ratio is probably due to, or derived from, the hornblende and biotite content of the amphibolite facies material. Kretz (1959) has shown that in coexisting metamorphic hornblendes and biotites Mg/Mg + Fe + Mn + Ti values for mineral pairs fall on a straight line of unit slope. Moxham (1965) lists distribution coefficients of 1.0 and 0.95 for Fe and Mg between coexisting hornblendes and biotites, this showing the constancy of Fe/Mg ratios in the two minerals. Thus the trend in amphibolite facies gneisses is satisfactorily explained by equilibration of Fe and Mg in hornblende and biotite during the amphibolite facies metamorphism. However the fact remains that the 7 hypersthene-gneiss analyses fall within the main trend, although containing ortho- and clinopyroxene. Mueller (1961) has shown that the two pyroxenes do not usually have equal Fe/Mg ratios, and thus their crystallisation at granulite facies would not affect any earlier trend of constant Fe/Mg. Therefore, as Sheraton (1970) has suggested from Lewisian gneisses at Drumbeg, there is a possibility that the observed trend, at least in the granulite facies gneisses, is a relic of equilibration of Fe and Mg in hornblende and/or biotite during pre-granulite facies, lower grade metamorphism. It is not considered that the trend, which parallels the normal calc-alkaline trend of igneous differentiation, reflects any magmatic origin for these rocks, but is a purely metamorphic feature, and perhaps common to gneissose rocks from many Shield terranes. Skinner (1970) reports similar trends, though not so marked from gneisses elsewhere in the Lewisian and East Greenland. The markedly higher MgO content of the diopside-gneisses is reflected in their composition field.

None of the gneisses show any relationship to the low temperature trough on an An-Ab-Or diagram (Fig. 13a). The nebulitic gneisses, which

show features consistent with their mobilisation during D_3, plot as a cluster centred on Or:Ab:An = 18:58:24. The other gneisses have a marked scatter indicating their variable contents of K_2O, Na_2O and CaO. The clustering of nebulitic gneisses indicates their homogeneity as a group, which is confirmed by their field appearance and petrographic examination. In view of the variability of the other gneisses, the homogeneity of the nebulitic gneisses may have a bearing on their origin. It possibly indicates that they were formed selectively, bands in the migmatites of their composition being most affected by the processes leading to their mobilisation. No evidence is seen on this diagram for their being derived by partial melting.

A plot of Niggli k against mg (Fig. 16a) shows the wide variation in K_2O in the gneisses except the nebulitic gneisses which cluster about k = 0.21. This graph also emphasises the difference of the diopside-gneisses, whose mean mg is 0.762 compared to 0.442 for the rest of the gneisses. Nebulitic gneisses have a significantly lower mg than migmatitic gneisses (0.392 compared with 0.463), which is due to the lower hornblende and biotite content of the nebulitic gneisses. Hypersthene-gneisses, migmatitic gneisses and epidote-chlorite-gneisses all have comparable mean mg's (0.435, 0.483 and 0.442 respectively - a low value of 0.225 in an epidote-chlorite-gneiss was neglected).

An examination of the norms of the various gneisses (see Appendix 3), excluding the diopside-gneisses which are established to be markedly different in major element composition, shows only one possibly significant difference. The hypersthene-gneisses are often diopside normative, whilst the migmatitic and nebulitic gneisses are not. Such a difference is not apparent from the diagrams previously discussed. It perhaps lies in the fact that although hypersthene-gneisses are more aluminous relative to the other gneisses, they are even more calcic (Table 3). Thus insufficient Al_2O_3 is present to account for all CaO in felspar in the norm calculation procedure, and in the absence of enough CO_2 to take the excess into calcite, it is combined with some MgO and FeO to give diopside. The CO_2 analyses of the hypersthene-gneisses are considered dubious and, in the absence of carbonates in thin sections, may be too high, so that all these rocks may be diopside normative. Direct comparison of the means in Table 3 shows the significantly more basic nature of the granulite facies gneisses.

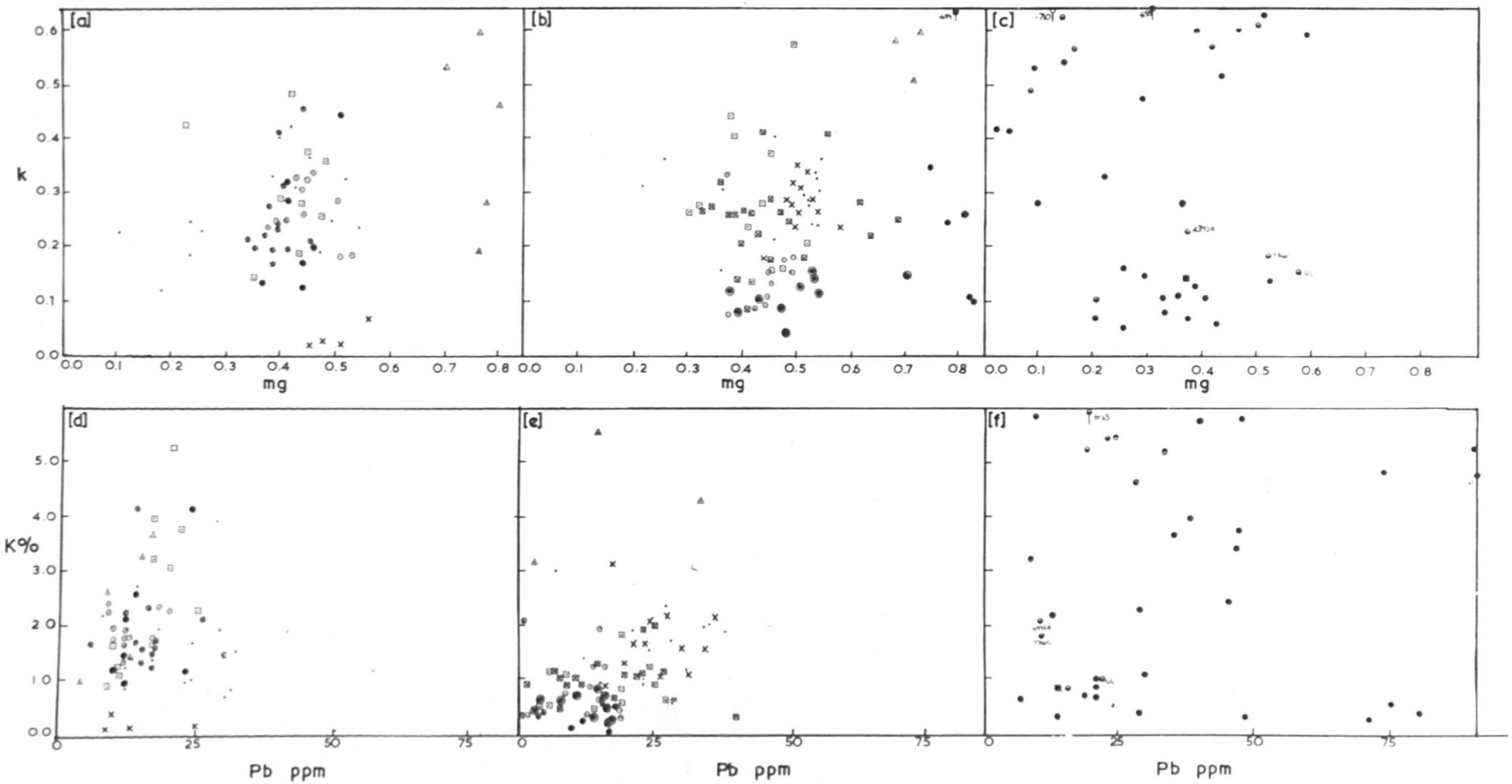


Fig.16: graphs of Niggli's k against mg and K against Pb

Chromium (Cr) abundances of all the hypersthene-nebulitic, epidote-chlorite- and diopside-gneisses are reasonably constant, averaging 25 ppm. Because of the homogeneous nature of all the samples of these rock types this is probably a good estimate of the mean value for the gneisses and is the same as the abundance level given for granodiorites by Taylor (1965). The migmatitic gneisses however have a higher mean value (91 ppm) and the Cr content is very variable. Inhomogeneous specimens of these petrographically variable rocks has probably caused this. The presence of thin hornblendic or biotite-rich bands in an otherwise acid gneiss will not significantly alter the major element abundances, but as both minerals concentrate Cr (and Ni) they will disproportionately elevate its abundance.

Taylor (1965) has suggested that Cr^{3+} should preferentially enter Fe^{3+} positions because of its close size to that cation and more ionic bond with oxygen, consequently being concentrated in early pyroxenes and being soon depleted in igneous fractionation. In the Skaergaard intrusion it shows very limited entry into olivines and ilmenite compared with pyroxenes and magnetite (Wager and Mitchell, 1951). From the standpoint of bonding energies Nockolds (1966) considered that Cr^{3+} could, by coupling as CrAl, substitute for $Fe^{2+}Si$ or MgSi.

Cr shows weak positive correlations with Fe_2O_3, FeO and MgO in all groups, and presumably its concentration is dependant on the proportions of mafic minerals. It can be contained in hornblendes and biotites as well as pyroxenes.

Low Cr/Ni ratios are present in the majority of samples (Fig.17). Moxham (1965) presents analyses of metamorphic hornblendes and biotites which show average Cr/Ni ratios of about 3.5 and 1.5 respectively. The low Cr/Ni ratios of the amphibolite facies gneisses then are due to their greater content of biotite. Now pyroxenes are expected to have much higher Cr/Ni ratios (Taylor 1965) yet the hypersthene-gneisses maintain the low value of the lower grade material. This would be consistent with a hypothesis of biotite- or hornblende-rich assemblages before granulite facies metamorphism, their Cr/Ni ratios being preserved on recrystallisation. The high K_2O content of the hypersthene-gneisses would favour the presence of biotite in hydrous assemblages.

Nickel (Ni), like Cr, abundances are the same as those found in granodiorites (average 18 ppm) except in the migmatitic gneisses (average 52 ppm).

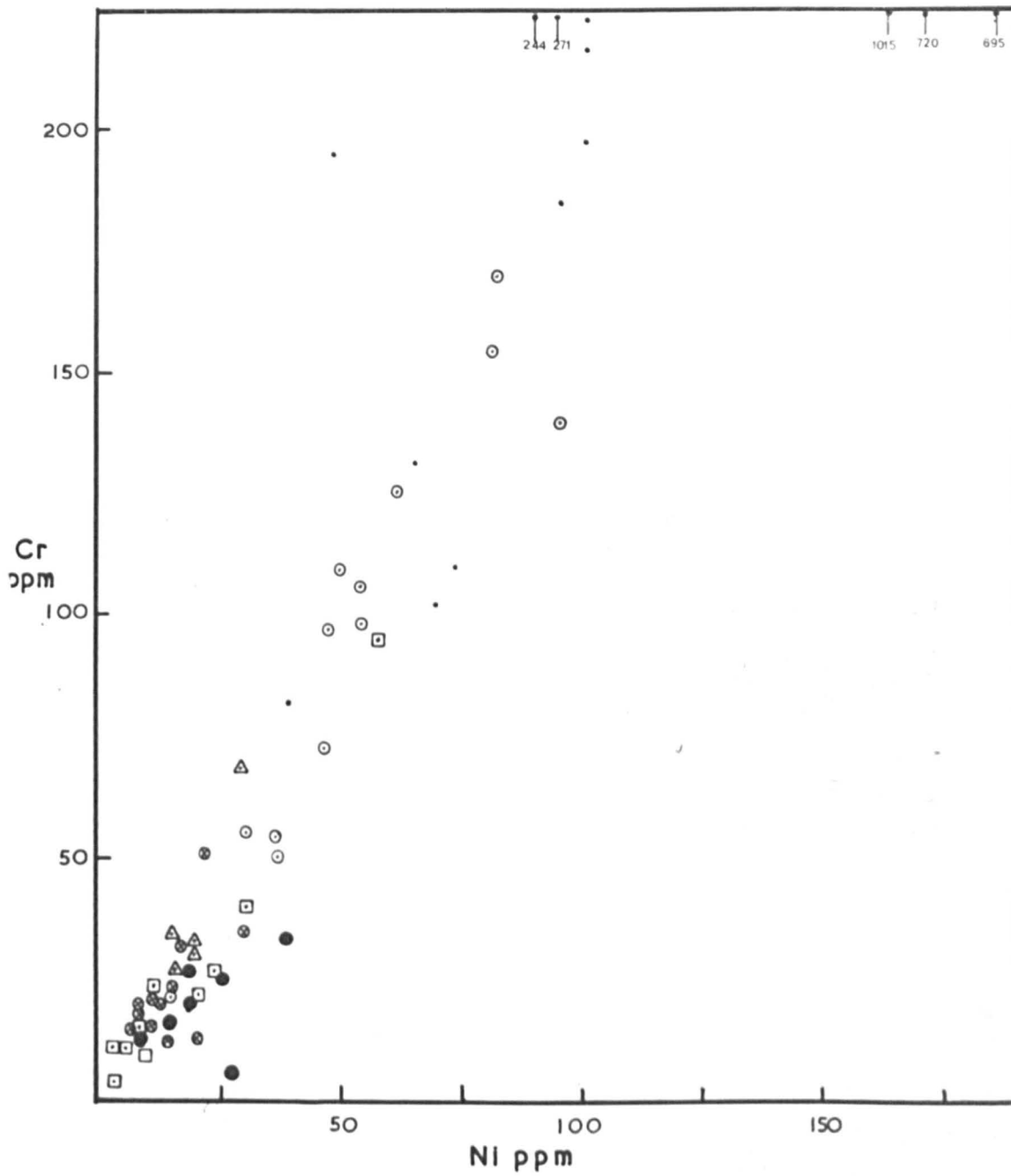


Fig.17: graph of Cr against Ni for quartzo-felspathic gneisses and garnet-biotite-gneisses. Symbols as Fig.13a.

chlorite-gneiss (288) relative to their parental migmatitic gneisses is probably due to the replacement of biotite by chlorite and the presence of Rb in felspars only.

The significance of the observed variations in K/Rb are returned to later.

Strontium (Sr) has an ionic radius intermediate between those of K and Ca. Taylor (1965) suggests that it is largely present in felspars. Sr^{2+} is too large to exhibit camouflage with Ca^{2+}. Sr may thus be expected to increase relative to Ca in felspars during igneous fractionation. Sr^{+} tends to be captured in early K^{+} positions, entering them more readily than Ca^{2+} so that the Ca/Sr ratio is lower in potash felspar than in plagioclase. The smaller size of $Sr^{2}+$ compared with K^{+} precludes their entry into 12-fold co-ordination in micas, substituting for K, and thus Sr is not present in micas to any extent. In contrast they are too big to enter Ca^{2+} positions in pyroxenes to any marked degree. Taylor (1965) suggests that Sr preferentially enters 8- or 10-fold co-ordination positions, the Ca sites in plagioclase, apatite and sphene, and K sites in potash felspar being about the right size.

The mean Sr content of the hypersthene-gneisses (894 ppm) is roughly twice those of migmatitic, nebulitic and epidote-chlorite-gneisses (327, 431 and 433 ppm). The diopside-gneiss is very low in Sr (105 ppm) compared with the other gneisses. Sr shows no significant correlations with CaO (Fig. 19a) or other elements in the hypersthene-gneisses, diopside-gneisses and epidote-chlorite-gneisses. In the migmatitic and nebulitic gneisses it correlates positively with Al_2O_3 and Na_2O, effectively with plagioclase, showing that it is contained mainly in that mineral. In the other gneisses it is probably distributed between plagioclase and alkali felspar, this possibly accounting for poor correlations.

Because of their high Sr and low Rb contents, the hypersthene-gneisses have very low Rb/Sr ratios (average 0.032) compared with the other gneisses. The hypersthene-gneisses' Rb/Sr ratios are much lower than that accepted for the crust (average 0.24 : Taylor 1964) and are more comparable with values obtained from oceanic tholeiites and chondritic meteorites. The range of Rb/Sr in the gneisses is shown in Fig. 25d. Its significance is discussed later.

TABLE 3 : Average compositions of major gneiss groups

	1	2	3	4	5	6
SiO2	59.33	62.81	67.45	60.81	64.56	63.95
TiO2	0.54	0.54	0.31	0.23	0.48	0.46
Al2O3	17.5	15.6	16.4	12.7	15.3	16.0
Fe2O3	3.02	2.75	1.64	1.51	2.65	2.46
FeO	2.53	2.67	1.33	2.04	1.48	2.00
MnO	0.07	0.06	0.04	0.16	0.05	0.06
MgO	2.3	2.5	1.0	6.4	1.6	1.8
CaO	4.88	3.46	3.31	8.14	3.29	3.61
Na2O	4.7	3.9	4.5	2.7	3.8	4.2
K2O	2.34	2.32	2.24	2.87	3.19	2.50
P2O5	0.28	0.11	0.09	0.08	0.16	0.14
H2O	0.96	1.52	0.67	1.76	1.44	0.94
CO2	0.24	0.45	0.46	0.41	0.42	0.41
Trace elements in parts per million						
Cr	23	91	22	38	25	45
Ni	17	52	15	21	17	29
Rb	30	100	67	72	92	78
Sr	894	326	446	105	433	477
Y	7	7	6	10	9	7
Zr	144	165	175	137	188	170
Nb	4	7	4	5	6	6
Ba	1209	543	722	542	927	791
La	32	22	31	17	48	32
Ce	65	40	61	39	82	60
Pb	15	13	17	11	16	15
Th	1	9	11	8	6	7
K/Rb	647	207	266	331	288	266
Ba/Rb	42.4	5.5	11.1	7.0	9.9	10.1
K/Ba	16.7	37.6	28.9	41.5	31.2	26.2
Rb/Sr	0.034	0.316	0.157	0.667	0.262	0.164
Ba/Sr	1.34	1.54	1.62	5.25	2.36	1.66
Ca/Y	7470	3750	3660	5690	2596	5157
Cr/Ni	1.3	1.9	1.5	1.8	1.5	1.6

1. Average of 7 analyses of hypersthene-gneiss.
2. Average of 14 analyses of migmatitic gneisses.
3. Average of 12 analyses of nebulitic gneisses.
4. Average of 5 analyses of diopside-gneisses.
5. Average of 10 analyses of epidote-chlorite gneiss.
6. Average of all gneisses except diopside-gneisses.

Yttrium (Y) has comparable abundances in all the gneisses (Table 3). However, the mean values are low compared with Taylor's (1965) estimate of the element's abundance in comparable crustal rocks (30 ppm for granodiorites).

Taylor (1965) states that Y^{3+} should be expected to be captured in early Ca^{2+} positions because of the similarity in size of the two cations. Y^{3+} is larger than Ca^{2+}. However it is found to concentrate in late pyroxenes, apatites and sphenes in the Skaergaard intrusion, but not in plagioclases. Cruft (1966) has shown that apatite concentrates Y up to about 1700 ppm, this being explained by its large 9-fold co-ordination site for calcium.

In the hypersthene-gneisses Y, together with La and Ce, correlates positively with FeO, MgO and to a lesser extent with CaO, indicating their concentration in the pyroxenes. The correlation of Y with FeO and MgO in migmatitic and nebulitic gneisses suggests that the element must be concentrated in hornblende or biotite. Strong positive correlations of Y with TiO_2, Fe_2O_3, CaO and P_2O_5 in epidote-chlorite-gneiss shows that it is contained in sphene epidote and apatite.

In igneous rocks Ca/Y is frequently seen to decrease with fractionation, due to the concentration of the larger Y ion in late stage fractions. although Y values in the gneisses are all similar, Ca/Y is much higher in the hypersthene-gneiss than in other types (Table 3).

Zirconium (Zr) forms a separate phase, zircon, and as its cation is not close in size to any common cation it rarely shows any marked substitution. It may show minor substitution for Ti^{4+} accompanying it in substitution for Fe^{3+} (Taylor 1965).

The abundance levels of Zr are similar in all types of gneiss, and its variability is low. It shows no significant correlations with any major element in these rocks.

Niobium (Nb) is present in low concentrations in all the gneisses, especially hypersthene-gneiss (mean 2 ppm) compared with a crustal estimate of 20 ppm (Taylor, 1965). In view of its tendency to form complexes such as $(NbO_4)^{3-}$ in volatile fluids (Taylor, 1965) its relative depletion in these Lewisian metamorphic rocks is hardly surprising.

Barium (Ba) substitutes for K^+ only amongst the common cations, and as expected shows a strong positive correlation with K_2O in the gneisses (Fig. 19a). In igneous K minerals Ba^{2+} behaves according to the classical capture principle, but although entering early K^+ sites is not depleted until late stages in magmatic differentiation (Taylor 1965). Nockolds (1966) suggests that Ba^{2+} might substitute for Ca^{2+} at high temperatures.

Ba has a much greater concentration in the hypersthene-gneisses (average 1209 ppm) than in migmatitic, nebulitic or diopside-gneisses (543, 685 and 541 ppm), a comparable content being present in epidote-chlorite-gneisses (927 ppm). Its relative concentration in the granulite facies rocks is expressed by their low K/Ba, and, high Ba/Rb ratios, compared with lower grade rocks (Table 3 Fig. 24d and 25a). Discussion of the variation of these two ratios is continued later in the chapter. Ba/Sr ratios are not significantly different in the hypersthene-, migmatitic, nebulitic and epidote-chlorite-gneisses, averaging 1.34, 1.54, 1.62 and 2.36 respectively. The low Sr of the diopside-gneisses gives a high Ba/Sr ratio of 5.25.

Rare earth elements lanthanum (La) and cerium (Ce) are complex geochemically. Although substitution is known to occur in most minerals, both felspars and mafic minerals, rare earths are often present in rare or accessory minerals such as monazite, allanite, zircon, sphene, epidote and apatite (Haskin et al., 1966). However the bulk of rare earths in a rock may be incorporated in major rock-forming minerals. Mafic minerals tend to favour heavy rare earths and have higher rare earth concentrations than felspars which favour the lighter elements (Haskin et al., 1966).

The contents of La and Ce are very variable in all gneiss groups, and there is no significant difference between the mean abundance values which are the same as crustal values. The Ce/La ratio is remarkably constant at about 2, which again is that found in crustal rocks (Fig. 23b). Correlation with FeO, MgO and CaO in hypersthene-gneisses shows the entry of Ce and La into pyroxenes in these rocks. In the amphibolite facies gneisses, they show strong coherence with P_2O_5, entering the commodious apatite structure where they may substitute for Ca^{2+} (Cruft, 1966). In the epidote-chlorite-gneisses Ce and La show very strong positive correlation with Zr, indicating entry into the zircon structure. They show a much stronger correlation with each other and with Th however, and may be present in another phase, perhaps allanite. In some epidote-chlorite-

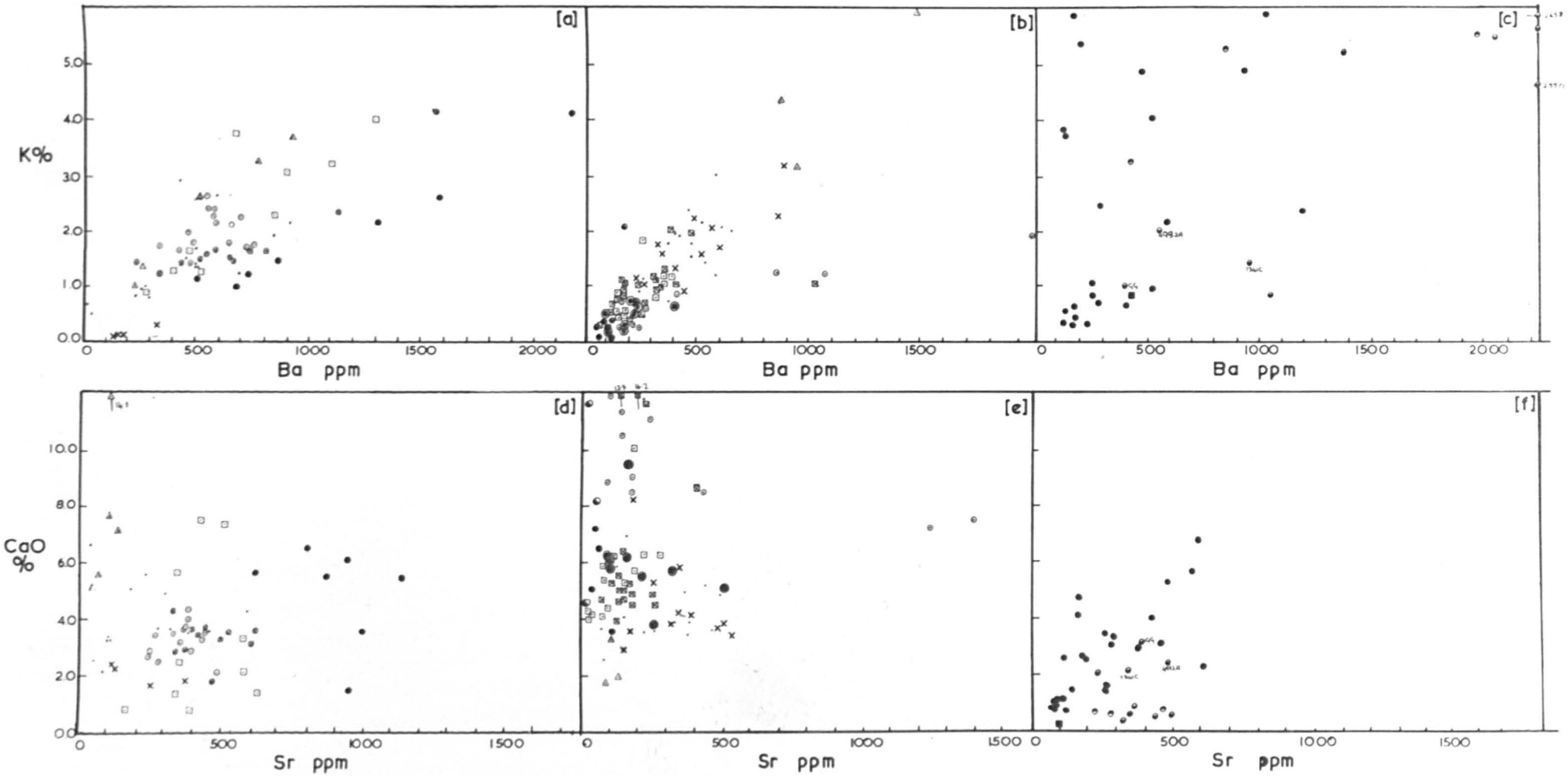


Fig.19: graphs of K against Ba and CaO against Sr.

gneisses, Ce/La ratios are less than 2 suggesting some enrichment in La relative to Ce (Fig. 23b) during greenschist facies metamorphism.

Lead (Pb), being intermediate in ionic radius between K^+ and Ca^{2+} might be expected to be present in K-felspar and micas, and in plagioclase apatite and other lattices with large Ca^{2+} sites (Taylor, 1965). To complicate matters Pb can be present as a sulphide phase also, and up to 20 per cent can be extracted from felspars in acid. In K-felspars K/Pb decreases with metamorphic grade, Pb substitution increasing with physical conditions (Heier, 1960). Nockolds (1966) states that as Pb^{2+} will enter a lattice less readily than K^+ and K/Pb should decrease in igneous fractionation.

In the gneisses, mean Pb abundances are normal and are the same in all types, as is their variability. Only in epidote-chlorite-gneiss does Pb show significant correlations, with K_2O, Al_2O_3 and Rb showing its presence in microcline. The mean K/Pb ratio for all types is the same, and there is little meaningful variation in the ratio (Fig. 16d).

Thorium (Th) is usually only found in rare and accessory minerals such as monazite, allanite, sphene and zircon (Taylor, 1965), but probably a large proportion in rocks is in a disperse intergranular phase, Th being readily leached from total rock samples.

In hypersthene-gneisses, Th is below the normal limits of detection of X.R.F. methods, actual values being less than 1 ppm or negative. The element is thus depleted in these rocks compared to normal crustal values of 10 ppm. In the other gneisses, more normal values have been obtained (Fig. 26a). In the diopside-gneisses and epidote-chlorite-gneisses Th correlates positively with Ce and La, and accompanies them in following TiO_2 and P_2O_5 indicating entry into sphene and apatite. Like Ce and La Th is probably also present in allanite in epidote-chlorite-gneisses.

The main differences between the groups of gneisses are shown below.

	Hypersthene-gneisses	Migmatitic gneisses	Nebulitic gneisses	Diopside gneisses	Epidote-chlorite-gneisses
Al_2O_3	17.5	15.6	16.5	12.7	15.3
CaO	4.88	3.46	3.29	8.14	3.29
K/Rb	647	207	266	331	288
K/Ba	16.7	37.6	28.9	41.5	31.2

	Hypersthene-gneisses	Migmatitic gneisses	Nebulitic gneisses	Diopside gneisses	Epidote-chlorite-gneisses
Rb/Sr	0.034	0.316	0.157	0.667	0.262
Ba/Rb	42.9	5.55	11.1	7.01	9.92
Ba/Sr	1.34	1.54	1.62	5.25	2.36
Ca/Y	7470	3750	3660	5690	2596

Considering the number of recent publications of average analyses of gneisses at different metamorphic grades from the Lewisian and other Shield areas, it is worth comparing the various mean analyses from Coll and Tiree with those of other areas. The diopside-gneisses, for which a special origin is envisaged (see below), are excluded from comparisons.

The bulk chemistry of the average gneiss (Table 3) compares favourably with those for the Canadian Shield (Fahrig and Eade, 1968) and SW Australia (Lambert and Heier, 1969) (Table 4). Comparison of trace element abundances with gneisses from SW Australia shows the gneisses under consideration to be comparable, except containing lower Pb and Th. When compared with Australian rocks from a similar metamorphic environment (i.e. Musgrave Range, SW Australia) even this difference is erased. Thus, considered as a whole the quartzo-felspathic gneisses of Coll and Tiree are comparable geochemically with most Shield gneisses.

Mean analyses of granulite facies hypersthene-gneisses are compared with the intermediate to high pressure granulite facies gneisses of the Musgrave Range SW Australia (Lambert and Heier, 1969) and Scourian gneisses from Assynt (Skinner 1970) in Table 4. The Tiree hypersthene-gneisses are more felspathic than the Australian rocks having higher Al_2O_3 and (Na_2O+ K_2O). Their much lower Rb and higher Sr contents are reflected in higher K/Rb, Ba/Rb ratios and lower Rb/Sr and Ba/Sr ratios. In comparison with the average for Assynt gneisses, the Tiree mean analysis shows lower SiO_2, higher total Fe and K_2O. Rb, Ba and Sr are all much higher than in the Assynt gneisses. However both series seem to be comparable in K/Rb, K/Ba, Rb/Sr, Ba/Rb, Ba/Sr and Ca/Y ratios. The highest K/Rb ratio (971) is not comparable with that from Assynt (> 2000) although the mean value (647) is so high compared with those granulite facies gneisses from the Musgrave Range (309) (Lambert and Heier, 1969) and rocks of comparable K_2O content in the Alps (409) (Sighinolfi, 1969), that it may be considered anomalously high even for granulite facies gneisses. That it is anomalous compared

TABLE 4 : Average compositions of crustal shield rocks

	1	2	3	4	5	6
SiO2	60.3	65.3	67.0	60.6	64.6	67.6
TiO2	1.0	0.53	0.5	0.9	0.47	0.33
Al2O3	15.6	15.9	14.5	15.4	15.7	15.2
Fe2O3	7.2	1.4	1.5	7.2	2.53	1.76
FeO		3.1	3.0		1.95	1.28
MnO	0.12	0.08	0.26	0.2	0.06	0.05
MgO	3.9	2.2	2.5	3.9	2.2	1.3
CaO	5.8	3.4	4.0	5.7	4.5	3.22
Na2O	3.2	3.9	2.5	2.8	4.6	4.2
K2O	2.5	2.87	3.0	2.6	1.15	2.58
P2O5	0.24	0.16	0.2		0.16	0.11
Trace elements in parts per million						
Cr	100	76	-	-	48	30
Ni	75	20	-	-	37	19
Rb	90	-	115	70	13	75
Sr	375	320	135	340	566	487
Y	30	-	-	-	7	5
Zr	165	-	220	310	196	183
Nb	20	-	-	-	-	-
Ba	425	650	610	1090	770	829
La	30	-	-	-	-	-
Ce	60	-	-	-	41	44
Pb	12.5	-	35	20	12	16
Th	10	-	20	2.1	5	7
K/Rb	232	-	217	309	730	280
Ba/Rb	4.72	-	5.3	15.6	59.2	11.5
K/Ba	46.2	39.1	40.8	19.8	12.0	25.8
Rb/Sr	0.24	-	0.85	0.21	0.02	0.11
Ba/Sr	1.13	2.04	4.52	3.21	1.36	1.70
Ca/Y	1258	-	-	-	4590	4598
Cr/Ni	1.3	3.8	-	-	1.29	1.5

1. Crustal average (Taylor, 1965).
2. Average composition of the Canadian Shield (Fahrig and Eade, 1968).
3. Average composition of the Archaean South-West Shield of Australia (Lambert and Heier, 1968).
4. Average composition of granulite facies gneisses of the Musgrave Range, Australia (Lambert and Heier, 1968).
5. Average composition of 154 Assynt gneisses (Skinner, 1970).
6. Average composition of 22 gneisses from North Harris and Lewis (Skinner, 1970).

to generally accepted limits for crustal values of K/Rb (150-300) is obvious from Fig. 18a and calls for special explanation for the depletion of Rb relative to K.

The average analysis for lower grade gneisses from Coll and Tiree, excluding the diopside gneiss, bears a strong similarity to that for 22 Lewisian gneisses from North Harris and Lewis, especially when the K/Rb, K/Ba, Ba/Rb, Rb/Sr, Ba/Sr and Ca/Y ratios are compared (Table 4). Like the Hebridean examples, which have been attributed to "late Scourian" gneissification (Myers 1970) and the Laxfordian (Skinner 1970), the composition of the amphibolite facies gneisses are comparable with estimates of the mean composition of the continental crust (Taylor, 1965) (Table 4).

Garnet-biotite-gneisses.

The mean composition of 14 analysed garnet-biotite-gneisses is given in Table 5, for comparison with analyses of various types of quartzo-felspathic gneiss in Table 3.

On an ACF diagram (Fig. 15a) most plot in the garnet-plagioclase-hornblende field of the amphibolite facies. Significantly, they cluster away from the field of quartzo-felspathic gneisses towards the F apex, showing their Fe- and Mg-rich nature.

Fe-Mg-Alk plots of garnet-biotite-gneisses (Fig. 14a) although partly continuous with the trend of the major gneiss groups, deviate towards the Fe apex. They are relatively poorer in alkalies than the other gneisses, and show a marked variation in Fe/Mg ratios, tending to be Fe-rich. This last feature is emphasised by a plot of al-alk against Fe/Mg (Fig. 20b) in which they are compared with granofelses.

Plots of garnet-biotite-gneisses for Niggli k against mg (Fig. 16a) shows that they are a variable group compared with the other gneisses. As mg decreases, in response to increasing Fe content, k also decreases. This would indicate that in iron-rich gneisses, garnet has formed rather than biotite, (garnet co-existing with biotite in these rocks has a much lower mg value - Table 2), k thereby not being held in the rock.

Norms of the garnet-biotite-gneisses frequently contain significant amounts of corundum, especially those containing Al_2SiO_5 polymorphs. Their aluminous nature and high Fe/Mg ratios results in their containing abundant garnet. Although containing less total Al_2O_3 than the other

TABLE 5 : Average compositions of metasedimentary rocks

	1	2	3	4	5
SiO2	59.4	51.5	60.7	64.4	60.91
TiO2	1.16	1.10	0.89	0.62	0.98
Al2O3	14.0	13.1	15.6	15.5	16.72
Fe2O3	4.32	6.67	3.07	6.54	1.21
FeO	7.25	7.57	2.57		6.80
MnO	0.17	0.19	0.08	0.08	0.09
MgO	3.3	7.7	4.3	3.12	3.28
CaO	3.95	2.78	4.07	2.22	3.37
Na2O	2.9	4.6	2.1	3.74	2.55
K2O	1.71	0.23	2.74	2.44	2.30
P2O5	0.14	0.17	0.16	-	0.16
H2O	1.33	1.90	-	-	-
CO2	0.57	1.02	-	-	-
Trace elements in parts per million					
Cr	314	125	188	-	
Ni	101	86	60	91	
Rb	60	9	67	88	
Sr	224	151	442	424	
Y	16	17	26	-	
Zr	125	151	281	196	
Nb	7	8	-	-	
Ba	426	192	1258	-	
La	41	32	-	-	
Ce	42	95	71	-	
Pb	21	14	21	-	
Th	6	3	7	-	
K/Rb	285	212	340	230	
Ba/Rb	7.1	21.3	13.5	-	
K/Ba	33.3	9.9	1.81	-	
Rb/Sr	0.421	0.060	0.151	0.207	
Ba/Sr	3.45	1.27	2.85	-	
Ca/Y	1765	1144	1119	-	
Cr/Ni	3.2	1.4	3.1	-	

1. Average of 14 analyses of garnet-biotite-gneiss : Coll and Tiree.
2. Average of 4 analyses of anthophyllite-gneiss : Coll and Tiree.
3. Average of 10 garnet-biotite-gneisses; South Harris, Skinner (1970).
4. Average composition of Precambrian greywackes, Wyoming (Condie, 1967).
5. Metasediment analysis by Misch (1968).

gneisses, they are significantly poorer in total $CaO+Na_2O+K_2O$, so that less Al_2O_3 is held in felspars, and more available for garnet.

Further significant differences from the quartzo-felspathic gneisses are the much higher TiO_2 and MnO contents of the garnet-biotite-gneisses. Mn is probably largely present in garnets (see Table 2).

Cr and Ni, largely contained in garnets and biotites (see Table 2), are variable. Ni lies within the same limits as in quartzo-felspathic gneisses, but Cr has a very wide variation, rising to a maximum of 1015 ppm in a very garnetiferous specimen. Many of the specimens have a similar Cr/Ni ratio to the gneisses (Fig. 17), that is about 2.0. Analysed biotite from specimen 4965B has a Cr/Ni ratio of 1.4, the coexisting garnet having a ratio of 3,6 (Table 2). The preferential entry of Cr into almandine garnet, where it presumably substitutes for Fe, leads to very high Cr/Ni ratios in the most garnetiferous specimens (Fig. 17). Cr shows strong positive correlations with Fe_2O_3, FeO and MnO as a result of its concentration in garnet. The average Cr/Ni ratio of 3.0 for the whole group is due to the inclusion of very garnetiferous specimens.

Rb contents and K/Rb ratios have similar ranges as in quartzo-felspathic gneisses (Fig. 18a) although K/Rb never reaches the high values as in hypersthene-gneisses. In view of the fact that biotite is the only mineral in these rocks containing large amounts of K and Rb, some of the K/Rb ratios are abnormally high. This may be the result of plagioclases with exceedingly high ratios or biotites with abnormally low Rb contents. The analysed biotite from a garnet-biotite-gneiss (Table 2) has a K/Rb ratio of 276

Sr, Zr, Ba and Th have lower abundances than in the other gneisses, and display wide, uncorrelated variations (for Ba and Sr see Figs. 19a and d). The variation, in Zr at least, may be accounted for by a sedimentary origin for garnet-biotite-gneisses, and due to variation in detrital heavy mineral fractions of the original rocks. In the case of Ba and Sr the result is wide variation in Ba/Rb, K/Ba, Rb/Sr, Ba/Sr and Ca/Sr ratios. A similar though not so pronounced, variation is seen in the abundances of Ce and La.

Y and Nb, which together with Ce and La are concentrated in garnets (see Table 2, and Haskin et al., 1966), are much higher in the garnet-

biotite-gneisses than in quartzo-felspathic gneisses. As a result, the garnetiferous gneisses show lower Ca/Y ratios than the other gneisses.

Pb, which is remarkable in its lack of correlation with any element, is in higher concentrations than in the quartzo-felspathic gneisses (Fig. 16d), and K/Pb ratios are lower.

Although not comparing favourably with the mean of 10 garnet-biotite-pelites from S. Harris (Skinner 1970) (Table 5), being more Ti- and Fe-rich and less aluminous, the garnet-biotite-gneisses do have some similar features. The Cr/Ni ratio of both has a common value of 3, Rb contents are comparable, although the Harris material has higher K/Rb and Ba/Rb ratios, both have equally high Pb contents and comparably low Ca/Y ratios. Although containing less Ba and Sr, the garnet-biotite-gneisses of Coll and Tiree have Ba/Sr ratios of the same magnitude as those from South Harris. It is worth noting that the South Harris material has been variably affected by alteration in the South Harris migmatite complex. Also included for comparison in Table 5 is a mean analysis of metasediments given by Misch (1968) and Condie's (1967) mean analysis of Precambrian grey wackes from Wyoming.

Anthophyllite-gneisses.

Examination of the mean analysis of 4 anthophyllite-gneisses (Table 5) and their fields of composition on various diagrams (Figs. 13a, 14a, 15a, and 16a) shows their contrast in composition to the quartzo-felspathic and garnet-biotite-gneisses. They are basic in composition, having high total Fe, and MgO contents, but very low CaO compared to their high alkali content. Sufficient Al_2O_3 is present to account for all the CaO in felspar, in fact, as mentioned in Chapter 3, the plagioclase has a sodic composition averaging An_{20}. As a result calciferous amphiboles are not present, most of the Fe and MgO being taken up in anthophyllite-gedrite. The low K_2O content (Fig. 16a) has led to only small quantities of biotite being present in the mode.

The low Rb content but normal K/Rb ratios (Fig. 18a) is probably caused by the small quantities of biotite, as this is the only mineral which can accept large amounts of the two elements. Sr, since it is largely precluded from

the biotite structure and cannot readily substitute for any ion in a non-calciferous amphibole, must be contained in the plagioclase. Ba, which closely follows K, is deplete compared with other rocks, as a result of the low K_2O content of the anthophyllite gneisses. However the very low K/Ba ratio (9.21), even when compared with granulite facies rocks, suggests that it is present in plagioclase as well as in biotite. Its positive correlation with CaO serves to confirm this. Phillpotts and Schnetzler (1970) report similarly low K/Ba ratios in igneous plagioclase phenocrysts. Like the garnet-biotite-gneisses, anthophyllite gneisses contain relatively high concentrations of Y, Nb, La and Ce, the analysis of an anthophyllite from these rocks showing them to be concentrated in this phase (Table 2). Pb which has a comparable concentration to the other rocks considered, correlates at a high degree with Na_2O, perhaps indicating its presence in plagioclase.

No comparable rocks from elsewhere in the Lewisian have been analysed, and a search for rocks of similar composition from elsewhere in the literature has proved fruitless. Rabbitt (1948) suggests that anthophyllite, apart from its presence in ultrabasic rocks, is a feature of aluminous or Ca-poor rocks of metamorphic terranes.

The main points arising from this section are:

1) Hypersthene-gneisses have lower Rb, Y, Nb and Th, and higher Al_2O_3, CaO, Sr and Ba than lower grade gneisses.

2) The metasedimentary rocks are geochemically distinct from the quartzo-felspathic gneisses.

3) The diopside-gneisses, although considered to be derived from quartzo-felspathic gneisses, are markedly distinct from them.

4) Most of the gneisses are comparable with Shield rocks from elsewhere, although the hypersthene-gneisses are abnormally poor in Rb, Y, Nb and Th. Garnet-biotite gneisses are comparable with South Harris metasediments.

5) There is some geochemical evidence of amphibolite facies assemblages having been present in the gneisses prior to granulite facies metamorphism.

Possible intrusive igneous rocks.

It was shown in earlier chapters that some metabasites and granofelses are in dyke-like bodies and have an igneous origin. A suggestion was made that the hornblende-peridotite ultramafic bodies were emplaced tectonically.

TABLE 6 : Average compositions of metabasites

	1	2	3	4	5
SiO2	49.7	49.7	49.8	49.6	
TiO2	1.41	1.53	1.42	1.34	
Al2O3	13.7	15.0	13.9	13.0	
Fe2O3	7.03	6.21	6.10	5.99	
FeO	8.58	7.79	9.64	8.30	
MnO	0.22	0.19	0.24	0.22	
MgO	6.6	6.9	6.3	6.6	
CaO	6.30	5.80	5.52	6.97	
Na2O	2.4	3.4	1.8	2.4	
K2O	1.00	0.60	1.05	1.16	
P2O5	0.11	0.11	0.11	0.12	
H2O	2.58	2.29	2.61	2.63	
CO2	0.69	0.54	0.73	0.74	
Trace elements in parts per million					
Cr	225	230	204	236	200
Ni	114	94	97	132	150
Cu	86	128	51	89	100
Zn	90	86	90	91	100
Rb	25	8	29	30	30
Sr	166	215	124	169	465
Y	23	21	23	23	25
Zr	89	89	87	91	150
Nb	7	5	5	9	20
Ba	260	208	237	297	250
La	10	10	9	11	11
Ce	18	20	16	24	35
Th	2	1	2	3	2.2
Pb	14	9	15	17	5
K/Rb	332	623	300	321	-
Ba/Rb	10.4	26.0	8.17	9.90	8.3
K/Ba	31.9	23.9	36.8	32.4	-
Rb/Sr	0.151	0.037	0.234	0.177	0.065
Ba/Sr	1.56	0.97	1.91	1.76	0.54
Ca/Y	1917	1972	1714	2164	-
Cr/Ni	1.97	2.50	2.10	1.79	1.33
Fe/Mg	2.61	2.58	2.07	2.95	-

1. Average composition of 41 metabasites.
2. Average composition of 9 pyribolites.
3. Average composition of 12 garnet-amphibolites.
4. Average composition of 20 amphibolites.
5. Average trace element abundances in basalts (Taylor, 1965).

Structural and petrographic evidence shows that all three groups of rocks were present before the granulite facies metamorphism.

Metabasites.

The mean analyses of pyribolites, garnet-amphibolites and amphibolites are given in Table 6 together with Taylor's (1965) estimate of trace element abundances in basaltic rocks. In Table 7 the average of all metabasites from Coll and Tiree is compared with those for Scourie dykes from Drumbeg (Sheraton, 1970), basites from Gairloch (Park, 1966), norites from South Harris (Skinner, 1970), tholeiitic flood basalts from the Deccan (Washington, 1922) and tholeiites from Hawaii (Wentworth and Winchell, 1947). All samples, except three olivine normative types, contain normative quartz and are therefore oversaturated. Most contain abundant orthopyroxene in the norm except for 5 (C1,GA,GB,5641, and 7029A) which have high normative diopside and are from agmatite environments. The marked lack of diopside in the norm is due to the metabasites' CaO-poor nature which is well shown by a plot of Niggli c against Niggli mg (Fig. 20c). Nepheline is found only in the norm of GA, an agmatised garnet-amphibolite. Its presence may reflect Na influx in the agmatisation process, or the complete saussuritisation of the felspar in the rock. Although the metabasites could be termed norites in terms of their norms (Yoder and Tilley, 1962), their nature could be largely due to the presence of abundant hornblende, which is generally hypersthene normative, and the loss of Ca and alkalies during metamorphism. Amphibole-rich basic rocks in many metamorphic terranes show quartz-tholeiite affinities (Park, 1966; van de Kamp, 1968; Sheraton, 1970).

Fe-Mg-Alk and ACF plots (Figs. 14b and 15b) of metabasites show a marked grouping, suggesting little significant variation in their major element chemistry. On the Fe-Mg-Alk diagram (Fig. 14b) the group shows no significant trend, unlike the Scourie dykes of Drumbeg which show a prominent linear trend of iron enrichment. Like the Scourie dykes they are magnesium rich.

On the ACF diagrams (Fig. 15b), all those pyribolites containing garnet fall within the plagioclase-garnet-prthopyroxene field of the granulite facies.(Turner, 1968), together with some that are garnet-free.

Two garnet-free pyribolites plot in the plagioclase-clinopyroxene-orthopyroxene field as they should on theoretical grounds. The garnet-amphibolites and amphibolites are indistinguishable, most falling in the plagioclase-garnet-hornblende field of the amphibolite facies perhaps showing that the presence of garnet in these rocks was controlled by physical conditions rather than composition. Orville (1969) indicates that any rock, whatever its origin, composed chiefly of hornblende with some plagioclase, will fall on the ideal hornblende-plagioclase tie-line in ACF diagrams. As the hornblende-plagioclase tie-line passes through the general field of basalt composition, any rock normally called an amphibolite will fall in that field and will approximate a basaltic composition. The metabasite field in this case plots at some distance from the tie-line being slightly more aluminous, and maintains a vague trend opposed to it.

According to De Waard (1965), garnet and orthopyroxene characterise aluminous granulite facies rocks rather than calcic. Green and Ringwood (1967) suggest that in undersaturated basic rocks garnet-orthopyroxene-clinopyroxene-plagioclase is the assemblage in the low pressure granulite facies, oversaturated rocks containing clinopyroxene-orthopyroxene-plagioclase. Only at high pressures do oversaturated rocks show garnet and clinopyroxene growth. Kanisawa (1969) showed that in the Abukuma metamorphic belt, garnet-amphibolites are characterised by higher Fe/Mg ratios than garnet-free types. Most of the metabasites are oversaturated and calcium-poor, thereby satisfying two criteria for the presence of garnet in some of them. Chemical controlling factors for the presence or absence of garnet in these rocks then, may be the availability of Al_2O_3 for garnet formation, and the Fe/Mg ratio. High alkali content in a basic rock would tend to give more plagioclase, biotite and hornblende, leaving less Al_2O_3 for garnet formation. Comparison of average analyses for garnet-amphibolites and amphibolites shows the former contain less alkalies and CaO and more $FeO+Fe_2O_3$ than the latter, both having comparable Al_2O_3 and MgO contents. The pyribolites have more Al_2O_3 than either, but vary in their alkali and CaO contents.

Fig. 20a shows a plot of Niggli al-alk against Fe/Mg for all the metabasites, al-alk being an estimate of the availability of Al_2O_3 not contained in felspars. From this, the presence of garnet in the granulite facies pyribolites does not seem to be controlled by chemistry. As the garnet-forming reaction (p. 82) has not gone to completion, it was prob-

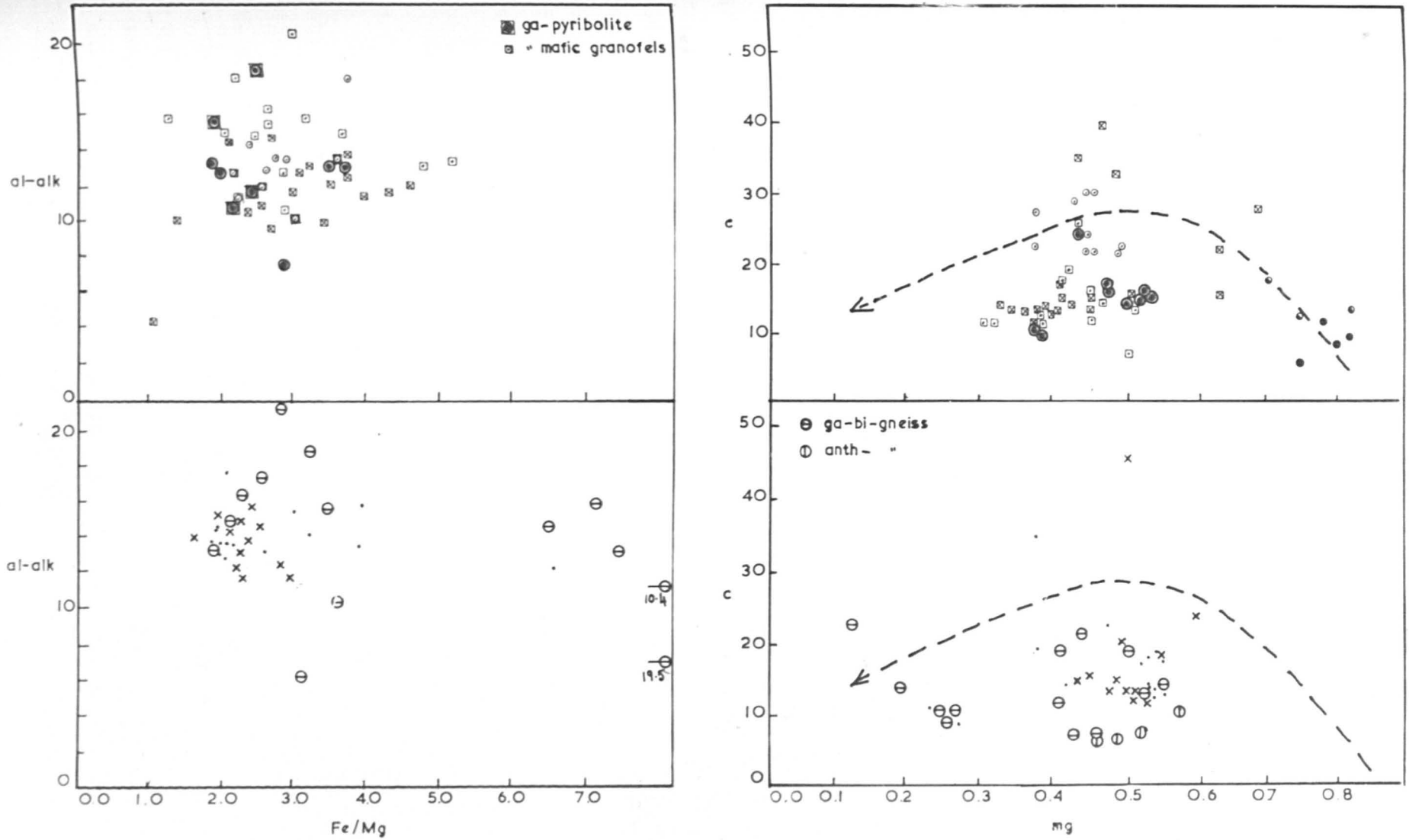


Fig.20: graphs of al - alk against Fe/Mg, and c against mg, for metabasites, granofelses and garnet-biotite-gneisses. Symbols as Fig.15b

ably more controlled by local chemistry and initiated by changing metamorphic conditions. The fact that these oversaturated rocks, initially garnet-free, developed garnetiferous assemblages is consistent with Green and Ringwood's (1967) hypothesis that garnet appears in oversaturated rocks when they are raised to high pressures in the granulite facies. The low Fe/Mg ratio of the garnet-pyribolites would tend to preclude the presence of garnet, on Green and Ringwood's account, at low pressures. In the largely recrystallised amphibolites, the garnetiferous types have a greater al-alk for a particular value of Fe/Mg than garnet-free types. This, coupled with their lower CaO content has given garnet-amphibolites more Al_2O_3 available for garnet formation on recrystallisation. In the amphibolites what available Al_2O_3 was released by the change in An content of plagioclase on retrogression must have been utilised wholly in the formation of hornblende and a little biotite. Notwithstanding strong evidence for chemical control of garnet crystallisation, some garnets were retrogressed to mixtures of hornblende, plagioclase and ore during the amphibolite facies (p. 81 and Plate 24a).

Leake (1964) suggests the use of plots of Niggli c against mg (Fig. 20a) to distinguish between para-amphibolites and those representing metamorphosed igneous differentiation series. The metabasites have a strong scatter devoid of meaningful trends, and tend to group at low values of c relative to the figured trend for Karroo dolerites. Whilst a basic igneous series might plot in a similar fashion, it is equally likely that the distribution reflects different proportions of hornblende and plagioclase as suggested by Orville (1969) and is a metamorphic feature.

TiO_2 positively correlates at high significance levels with P_2O_5 in all the metabasite groups. Both show negative correlations with MgO and CaO, and positive correlations with total Fe. Leake (1964) has suggested that negative relationships of Niggli's ti and p to Niggli mg are strongly indicative of amphibolites belonging to a differentiated igneous series. In differentiation of tholeiites, total Fe, TiO_2 and P_2O_5 are expected to increase with degree of fractionation.

An attempt to estimate the alkalinity of the metabasites, if they do represent an igneous series, by plotting Na_2O+K_2O against SiO_2 (Fig. 21a) was frustrated by a strong scatter. The variation probably represents the varying proportions of hornblende and composition of felspars. It could also be due to alkali loss or gain during metamorphism. A plot of

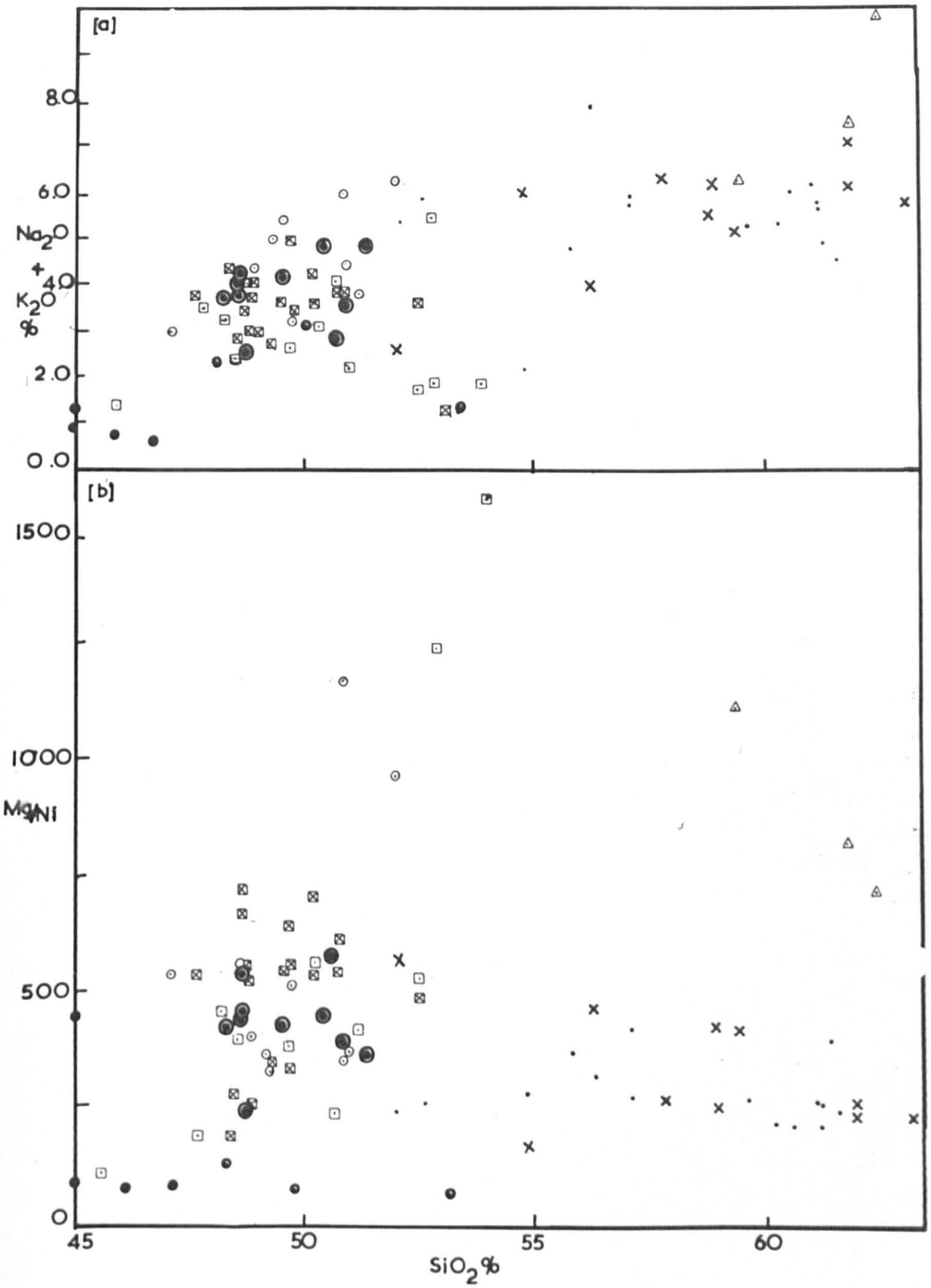


Fig.21: graphs of $Na_2O + K_2O$, and MgO/Ni against SiO_2 for granofelses and metabasites. Symbols as Fig.13b.

Niggli k against mg (Fig. 16) shows the pyribolites to be markedly lower in K_2O than the other metabasites.

Both Cr and Ni show strong positive correlations with MgO (Fig. 22a) in the metabasites. Leake (1964) has shown that in para-amphibolites formed from mixtures of pelites and dolomite a negative correlation of Cr and Ni with Niggli mg is expected as dolomites are poor in the elements and pelites richer. In contrast differentiation of basic magmas tends to concentrate Cr and Ni with MgO in early pyroxene and olivine crystals, so that all three decrease in later fractions. Cr and Ni also exhibit strong coherence with each other (Fig. 22c) which Leake has suggested to be more typical of meta-igneous rocks than sedimentary mixtures. As TiO_2 correlates negatively with MgO in the metabasites, it does likewise with Cr and Ni. Graphs of Cr and Ni against TiO_2 (Fig. 22b) for the metabasites closely follow the trends for igneous differentiates suggested by Leake (1964) who has shown that pelite-dolomite mixtures would show positive correlations. Now Cr and Ni must be largely contained in the mafic metamorphic minerals pyroxenes, amphiboles and garnets, in the metabasites, and their positive correlation with MgO depends on the varying abundance of the mafic minerals to a large extent, which itself is probably due to metamorphic banding and thus non-representative sampling. The overall Cr/Ni ratio in the metabasites of about 2.0 (Fig. 22c) is the same as that for a partially analysed brown hornblende from a pyribolite (78SW5B) and one from a garnet-amphibolite (C1) (see Table 2) and probably reflects the dominantly hornblendic nature of most of the specimens. A plot of MgO/Ni against SiO_2 for the metabasites (Fig. 21b) shows the tendency for this ratio to increase with SiO_2 content, a generally accepted feature of basic igneous series because of the preference for Ni^{++} rather than Mg^{++} in early igneous crystals. A simple increase in SiO_2 because of variation in mafic mineral content in the metabasites would not be expected to affect Mg/Ni, as this ratio would be reasonably constant in hornblendes stable under amphibolite facies conditions.

Rb is much lower in pyribolites than in garnet-amphibolites and amphibolites, and even though the former are low in K_2O, they have extremely high K/Rb ratios (average 623) compared with the amphibolite facies rocks (300

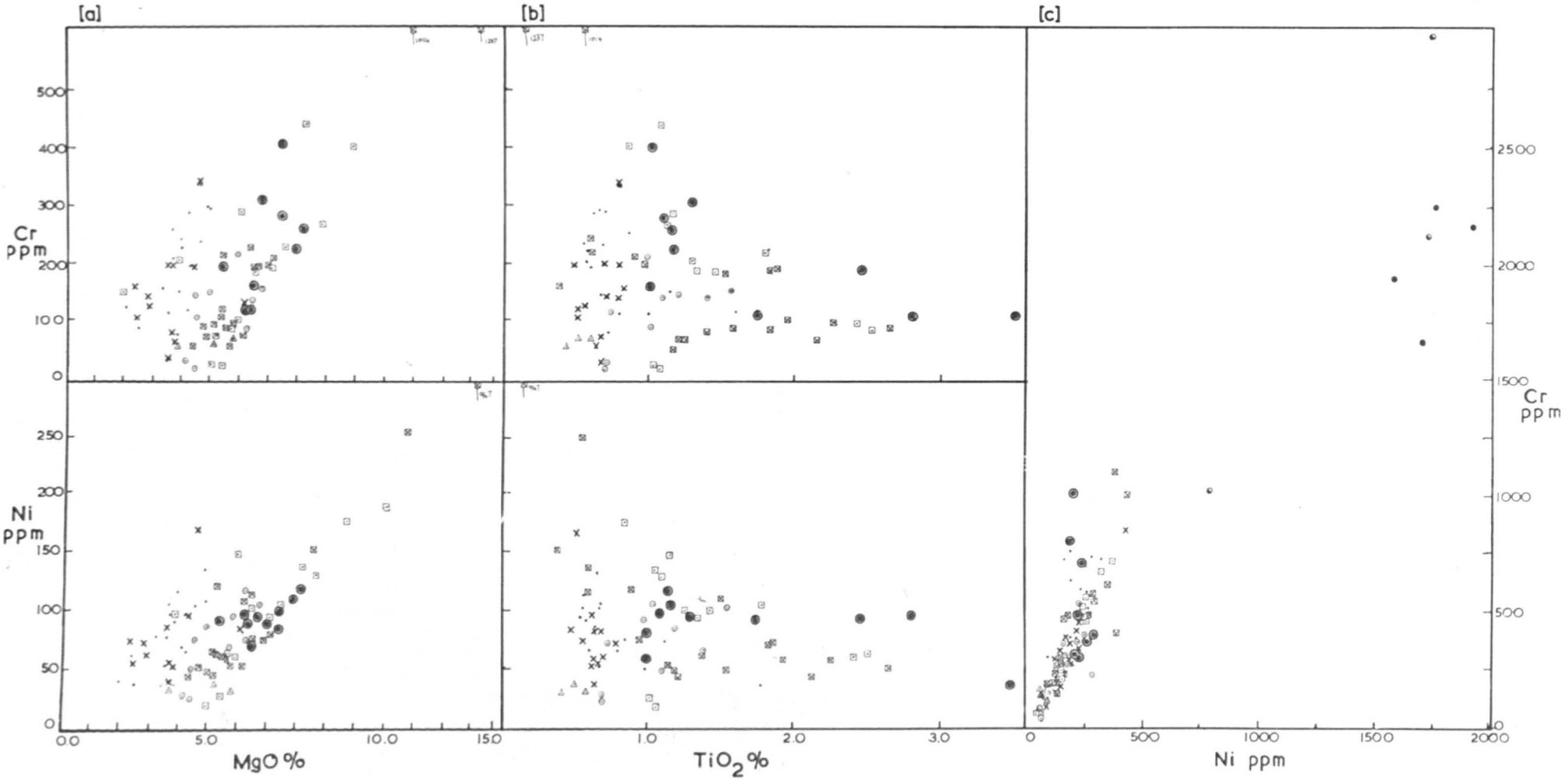


Fig. 22: graphs of Cr and Ni against MgO and TiO_2 and Cr against Ni for granofelses metabasites and ultramafic rocks.

and 321 respectively) (Fig. 18b). Analysis of a brown hornblende from a pyribolite (Table 2) and its co-existing plagioclase shows that Rb is contained in both, although in very low quantities in the fairly K_2O-rich hornblende, giving it an extremely high K/Rb ratio of 2426. In both garnet-amphibolites and amphibolites Rb correlates positively with K_2O, substituting for it in hornblende and plagioclase. K/Rb for a hornblende and plagioclase C1 are 355 and 320, similar to that for the rock itself.

Sr is low in the pyribolite hornblende and clinopyroxene but is strongly concentrated in the felspar, probably substituting for Ca^{2+}. Positive correlations with Na_2O and Al_2O_3 in amphibolites suggests its concentration in plagioclase in them too. Fig. 19e shows its low correlation with CaO.

Y, Zr, Nb, La and Ce show strong positive correlation with TiO_2, P_2O_5 and Al_2O_3 in all metabasites, indicating that they may be concentrated in sphene, ilmenite and apatite as well as in major ferromagnesian minerals. This could equally be due to the coherence of the three major elements, the trace elements only needing to be correlated with one to vary in sympathy with the others. The concentration of these elements in mafic minerals is demonstrated by analyses of a clinopyroxene, hornblendes and garnet from metabasites (Table 2).

Ba shows no strong association except with K (Fig. 19b), and data from hornblendes and felspars in Table 2 shows its concentration in both. It is also present in appreciable quantities in garnets. Its presence in a clinopyroxene from a pyribolite suggests that it may substitute for Ca^{2+} in this mineral, perhaps, as Nockolds (1966) suggested, at high temperatures.

Pb's correlation with Al_2O_3 and Na_2O indicates its presence in plagioclase, but it may also be in a sulphide phase. Pb shows no correlation with K (Fig. 16e).

The mean analysis for the metabasites shows some similarity to tholeiitic lavas from the Deccan and Hawaii (Table 7), but is lower in CaO, more iron rich and has a higher Fe_2O_3) FeO ratio. Compared with abundance data for trace elements in basic igneous rocks (Taylor 1965), the metabasites as a whole have normal Cr, Ni, Cu, Zn, Rb, Y, Ba, La, Ce and Th abundances and

TABLE 7 : Average compositions of some basic rocks

	1	2	3	4	5
SiO2	50.61	50.45	49.66	49.0	52.9
TiO2	1.91	2.33	1.51	1.4	0.70
Al2O3	13.58	14.94	12.58	13.9	14.8
Fe2O3	3.19	3.38	4.89	4.4	3.20
FeO	9.92	7.55	9.31	10.2	4.74
MnO	0.16	0.08	0.19	0.16	0.12
MgO	5.46	7.67	6.24	6.0	6.9
CaO	9.45	9.17	10.03	8.9	8.00
Na2O	2.60	2.84	2.08	2.9	3.5
K2O	0.72	0.35	0.65	0.9	1.74
P2O5	0.39	0.27	0.19	0.13	0.40
H2O	2.13	0.96	1.63	2.0	-
CO2	-	-	0.47		-
Trace elements in parts per million					
Cr			176	111	296
Ni			178	75	65
Rb			10	21	13
Sr			196	243	933
Y			23	41	13
Zr			116	136	85
Ba			307	284	1365
Ce			35	-	40
Pb			11	-	12
Th			3	-	5
Cr/Ni			0.99	1.48	4.55
Fe/Mg	24	14	2.3	2.4	1.2

1. Average analysis of 11 tholeiitic flood basalts: Deccan (Washington, 1922).
2. Average analysis of 10 tholeiites: Hawaii (Wentworth and Winchell, 1947).
3. Average analysis of 15 Scourie dykes: Drumbeg (Sheraton, 1970).
4. Average analysis of 19 basites: Gairloch (Park, 1966).
5. Average analysis of 2 norites: South Harris Igneous Complex (Skinner, 1970).

low Sr, Zr and Nb and high Pb. The pyribolites are depleted in Rb compared with crustal values.

A comparison of the metabasites with other Lewisian basic rocks shows that they are poorer in CaO, and richer in Fe_2O_3 and K_2O than Scourie dykes from Drumbeg and basites from Gairloch. They have comparable trace element abundances but are characterised by higher Cr/Ni ratios. In contrast to the pyroxene-granulites of South Harris (Dearnley 1963), which have been called norites and linked with the South Harris Igneous Complex by Skinner (1970), they are poorer in SiO_2 and CaO and much more iron-rich. These norites contain Sr and Ba in amounts far in excess of the Coll and Tiree metabasites, and also have higher Cr/Ni ratios.

Granofelses.

The mean analyses for the groups of granofelses are contained in Table 8 together with an average for circum-Pacific calc-alkaline andesites (Taylor and White, 1965), and dacites from the Canadian Shield (Goodwin, 1968).

From Table 8 it is obvious that the mafic granofelses show stronger affinities to the metabasites than to the other granofelses, although differing from them in several important respects. They are lower in total iron and MgO and much higher in CaO compared with the metabasites (Figs. 13b and 20c) and compare better with the South Harris norites in major element composition. Sr, Ba, La and Ce are all higher in the mafic granofelses than in metabasites, but Ba and Sr tend to be lower than in the South Harris norites. Various plots of mafic granofelses fall within the same fields as the metabasites, the correlation patterns of trace elements with major elements give rise to similar conclusions regarding their mineralogical control, and thus they may be grouped with the metabasites in the absence of any evidence for different origins. Like the pyribolites, although poor in K_2O, the mafic granofelses have exceptionally high K/Rb ratios (Fig. 18b) averaging 971.

Nearly all the garnet-biotite- and hornblende-biotite-granofelses are quartz normative up to 25 per cent. CaO is too low to make any of them diopside normative and thus only quartz, felspars and hypersthene appear

TABLE 8 : Average compositions of granofelses

	1	2	3	4	5	6	7
SiO2	49.8	58.2	60.1	61.2	59.0	60.1	61.5
TiO2	1.30	0.78	0.63	0.52	0.71	0.7	0.63
Al203	14.3	14.9	14.7	15.5	14.8	17.2	15.7
Fe203	4.19	3.37	3.22	1.98	3.30	6.1	1.83
FeO	7.89	5.28	3.38	1.7	4.44		4.49
MnO	0.19	0.15	0.11	0.06	0.13	-	0.16
MgO	5.3	3.8	3.6	4.9	3.7	3.5	2.38
CaO	9,52	4,91	5.58	2.39	5.21	7.1	4.41
Na20	3.4	3.3	3.4	2.3	3.3	3.3	3.15
K20	0.99	2.02	2.08	5.63	2.05	1.3	1.16
P205	0.19	0.15	0.15	0.61	0.15	-	0.12
H20	1.75	2.14	2.05	1.66	2.10	-	2.27
CO2	0.44	0.47	0.53	0.29	0.50	-	2.18
Trace elements in parts per million							
Cr	115	196	146	66	174	60-80	
Ni	71	85	78	33	82	25-40	
Cu	26	61	-	-	-	40-70	
Zn	91	98	-	-	-		
Rb	14	82	98	138	89	25-45	
Sr	406	273	317	112	293	220-320	
Y	18	20	17	10	19	25	
Zr	77	141	179	300	158	95-150	
Nb	4	6	6	8	6		
Ba	548	465	488	1122	475	230-350	
La	16	17	20	52	18		
Ce	37	36	45	94	40		
Pb	13	30	25	17	28	9-14	
Th	2	3	5	15	4	25-55	
K/Rb	586	204	176	338	191	240-330	
Ba/Rb	39.1	5.67	4.98	8.13	5.34		
K/Ba	20.9	36.1	35.4	41.7	35.8		
Rb/Sr	0.034	0.300	0.309	1.23	0.303		
Ba/Sr	1.35	1.70	1.54	10.0	1.62		
Ca/Y	3776	1753	2344	1706	1957		
Cr/Ni	1.62	2.31	1.87	2.00	2.12		

1. Average composition of 10 mafic granofelses : Tiree.
2. Average composition of 15 garnet-biotite-granofelses : Coll and Tiree.
3. Average composition of 12 hornblende-biotite-granofelses : Coll and Tiree.
4. Average composition of 3 diopside-granofelses : Coll.
5. Average compposition of 30 intermediate granofelses : Coll and Tiree.
6. Average andesite composition and trace element ranges in andesites (Taylor, 1966).
7. Mean of 272 analyses of Canadian Shield dac ites (Goodwin, 1968).

in most of the norms. Although the presence of corundum in the norm is very sensitive to small errors in CO_2 determination and hence apportion of CaO with Al_2O_3 to felspars, significant amounts of normative corundum are present in some granofelses' norms.

Evidence has been presented in Chapter 3 to suggest that the granofelses had an igneous origin, perhaps contemporaneous with the metabasites. However, in view of their chemical dissimilarity to the mafic granofels (the only definitely intrusive representatives), and mineralogical similarity to garnet-biotite-gneisses, a sedimentary origin for the garnet-biotite- and hornblende-biotite-granofelses is equally possible. Comparison of mean analyses for both types of granofels shows that they are, for most purposes, identical in major and trace element composition.

On an Fe-Mg-Alk diagram (Fig. 14b) granofelses plot as a scattered group with some degree of overlap with metabasites. Together, the two groups define a roughly calc-alkaline trend. In contrast, the garnet-biotite-gneisses show a wide scatter, trending at variance to the granofelses. The constancy of Fe/Mg in the granofelses compared with the garnet-biotite-gneisses is shown by a plot of al-alk against Fe/Mg (Fig. 20b).

Plots of c against mg (Fig. 20d) show that the intermediate granofelses cluster well with the metabasites, although showing no significant trends. Although some garnet-biotite-gneisses fall in the same field, they generally show a high degree of scatter.

On a graph of Na_2O+K_2O against SiO_2 (Fig. 21a) the granofelses, together with the metabasites show a vague trend of slowly increasing alkalies similar to calc-alkaline igneous trends.

Plots of granofelses on graphs of Cr and Ni against MgO and TiO_2 (Figs. 22a and b) maintain the correlations shown by the metabasites, but add greatly to the degree of scatter. In the plots of Cr and Ni against MgO, the scatter is a result of granofels trends at lower values of MgO/Cr and MgO/Ni, perhaps as a result of increased Cr and Ni substitution for Fe in the granofelses. Metasediments show no correlation in these four graphs and are not plotted.

In contrast to the trend of increasing MgO/Ni with SiO_2 in metabasites, the granofelses maintain a remarkably constant MgO/Ni ratio of about 250, possibly a result of mineralogical control (Fig. 21b). Metasediments show strong random variations in this ratio, and in MgO/Cr ratios to a more marked extent.

A metasedimentary series would be expected to show some degree of variation in those elements normally concentrated in heavy minerals, such as Zr, La, Ce, Y, Nb and Th. As well as showing high relative standard deviations for these elements, garnet-biotite-gneisses also vary a great deal in Sr and Ba. In contrast the granofelses are a relatively homogeneous group with respect to all these elements.

To a large extent the mineralogy of the intermediate granofelses explains their trace element abundances. Means of analysed garnet, biotite and quartz-plagioclase fractions of granofelses are given in Table 2. The garnet concentrates Cr, Ni, Y, La and Ce, and contains a fair proportion of Ba. Biotite contains high Rb, Ba and Nb, and large quantities of Y, Ce and La. Both contain significant quantities of Th. The plagioclase felspars concentrate Sr in particular and also Pb to a certain extent.

The low K/Rb and high Rb/Sr ratios of these rocks is a reflection of their prominent content of biotite. The concentration of Y and rare earths in garnets and biotites explains the high average content of these elements and the low Ca/Y ratios.

The diopside-granofelses, like the diopside-gneisses have Sr depleted relative to Rb and Ba, and have high Rb/Sr and Ba/Sr ratios.

Compared with Taylor and White's (1966) mean analysis for andesites from orogenic belts (Table 8) that for intermediate granofelses (excluding diopside-granofelses) is somewhat depleted in Al_2O_3 and CaO, and high in K_2O. Other major elements compare very well. Trace element abundances in the granofelses mainly fall within or close to the limits of variation of andesites. Cr, Ni, Rb and Pb are all slightly high. K/Rb ratios for the granofelses (200) are lower than those given for andesites (240) and the Rb/Sr ratios are much higher. Both are satisfactorily explained by the granofelses' large biotite contents, a result of the rocks' high K_2O content. The biotite is a metamorphic feature and may be the result of K enrichment during the amphibolite facies, Rb being slightly favoured in competition for K sites and Sr largely excluded from the biotite lattice. In view of the granofelses' content of large volumes of concordant, possibly anatectic, granite veins, their present compositions might well be more basic than originally and lower in SiO_2, Al_2O_3, CaO and Na_2O.

Compared with the mean major element composition of dacites from the Canadian Shield (Goodwin, 1968), apart from being more basic, the granofelses

show strong similarities. As has been mentioned, in the absence of possibly anatectic concordant granitic veins, the granofelses may have had a more acid initial composition.

Ultramafic masses.

In Table 9 the average of four analysed fresh hornblende-peridotites is compared with analyses of three altered examples successively further from the core of the large body at Bousd (see Fig. 10), and with the mean of six analyses of fresh material from the Lewisian ultramafic body at Drumbeg (Sheraton, 1970). Estimated trace element abundances in alpine-type ultramafic masses as given by Goles (1967) are also contained in this table.

The main major element features of the Coll and Tiree rocks are shown by plots of Niggli c against mg (Fig. 20c), Na_2O+K_2O against SiO_2 (Fig. 21a), an ACF diagram (Fig. 13b) and an Fe-Mg-Alk diagram (Fig. 15b). Like the Drumbeg material, which typifies Lewisian ultramafic masses (see Bowes et al., 1964), that from Coll and Tiree has low Al_2O_3, total alkalies, and Fe/Mg ratios. High oxidation ratios in their iron content is probably due to the alteration of olivines to serpentine and magnetic granules. There is no marked difference in major element chemistry between the Coll and Tiree rocks and those from Drumbeg. Possible important differences are the slightly higher Al_2O_3 and K_2O, and lower MgO of the material studied here. Comparison of trace element abundances shows no significant differences. However, because of their high average K_2O content the Coll and Tiree rocks have higher K/Rb ratios than those at Drumbeg.

Compared with typical alpine peridotites, which they resemble in structure, the ultramafic rocks are poorer in MgO, and richer in CaO, Al_2O_3, SiO_2, total Fe and alkalies.

The mean values for Cr and Ni in the hornblende-peridotites compare well with average values given by Goles (1967) for ultramafic rocks, but the Cr/Ni ratios are lower than expected (Fig. 22c), possibly due to the presence of much olivine in the rocks. Olivine, with its divalent sites for cations is expected to concentrate Ni relative to Cr, in contrast to pyroxenes whose trivalent sites accommodate Cr more readily.

Rb, averaging 7 ppm, is in rather large amounts for ultramafic rocks, compared with the range of 0.03 to 1 ppm suggested by Murthy and Stueber

TABLE 9 : Average compositions of ultramafic rocks

	1	2	3	4	5	6
SiO2	44.2	42.4	48.3	53.7	49.6	
TiO2	0.48	0.32	0.77	0.17	0.26	
Al2O3	5.3	6.1	9.7	5.5	7.0	
Fe2O3	5.00	6.77	2.76	2.60	2.67	
FeO	7.13	5.21	8.16	5.66	7.04	
MnO	0.21	0.17	0.19	0.20	0.28	
MgO	24.9	26.5	14.6	19.6	21.7	
CaO	5.59	6.14	8.17	11.90	4.42	
Na2O	0.59	0.66	1.84	0.74	0.42	
K2O	0.26	0.14	0.40	0.37	2.58	
P2O5	0.04	0.014	0.068	0.016	0.007	
H2O	3.22	4.40	3.18	3.16	3.10	
CO2	0.50	0.39	1.21	1.00	0.81	
			99.4	104.6	100.2	
Trace elements in parts per million						
Cr	1722	2336	1018	3040	2167	2400
Ni	1424	1783	662	1733	1715	1500
Rb	7	6	5	27	180	1
Sr	61	54	42	19	11	20
Y	7	6	8	3	14	5
Zr	33	24	58	17	19	1
Nb	2	1	1	1	4	1
Ba	122	94	74	68	189	0.4
La	1	1	1	1	1	3.3
Ce	1	1	1	1	1	1
Pb	14	9	5	4	1	50 ppb.
Th	1	2	1	1	1	1
K/Rb	308	192	664	114	119	-
Ba/Rb	17.4	15.7	14.8	2.5	1.05	0.4
K/Ba	17.6	12.4	44.9	45.1	113	-
Rb/Sr	0.114	0.111	0.119	1.42	16.4	0.050
Ba/Sr	2.00	1.74	1.76	3.58	17.2	0.02
Ca/Y	5702	7307	7291	28322	2257	-
Cr/Ni	1.21	1.31	1.54	1.75	1.26	1.6

1. Average composition of 4 fresh ultramafic rocks : Coll and Tiree.
2. Average composition of 6 fresh ultramafic rocks: Drumbeg (Sheraton, 1970).
3. Analysis of hornblende-orthopyroxene-plagioclase rim : Bousd, Coll.
4. Analysis of hornblendite rim : Bousd, Coll.
5. Analysis of biotite-tremolite rim : Bousd, Coll.
6. Trace element abundances in alpine-type ultramafic rocks (Goles, 1967).

(1967) for alpine peridotites. Similarly, comparing the mean values given here for the related elements Sr, Ba and Pb (61, 122 and 14 ppm) with estimated abundances in ultramafic rocks (.20, 0.4, and 0.05 ppm) given by Goles (1967), these elements are relatively strongly concentrated in the rocks being considered. Rb^{+}, Sr^{2+}, Ba^{2+} and Pb^{2+} are able to substitute for K^{+} because of their similar size. Compared with the latest estimate of K_2O for ultramafic rocks of 10-240 ppm (Murthy and Stueber, 1967) the mean value here of 0.26 per cent seems abnormally high, and sufficient to explain the abundance of these trace elements. Compared with Goles' (1967) estimate of 1040 ppm Na_2O in ultramafic rocks, the value here of 0.59 per cent is again very high. This abnormally high content of alkalies and related elements in Lewisian ultramafic rocks is probably due to the prominence of hornblende, whose crystal lattice allows much substitution of alkalies and ions of similar size. There are several possibilities which might account for the presence of large amounts of hornblende and the consequent effect on the geochemistry of the Coll and Tiree rocks. The hornblende might be of primary origin. It might have crystallised from a lherzolitic (two pyroxenes + olivine) rock with included fluids stable at deeper levels (O'Hara, 1967a) but causing hydroxylation of pyroxenes on introduction of the system into a regime of lower P-T conditions. It may have crystallised during granulite facies metamorphism, an anhydrous original lherzolitic rock, taking in some of the H_2O released by reactions in country rock enabling pyroxene hydroxylation to occur. Finally it may be due to amphibolite facies hydration and metasomatism of the ultramafic masses. The first case, and perhaps the second, point to the abnormal geochemistry of these rocks being an original property, with interesting implications. The last two account for the features simply by metasomatism of the rocks.

The hornblende in the fresh rocks has been shown to be earlier than that found in the surrounding granulite facies rim. The high K/Rb ratios of two of the analysed hornblende-peridotites (763 and 805) is not consistent with introduction of alkalies during amphibolite facies metamorphism which has imparted very low ratios to biotite rims to ultramafic bodies (Fig. 24b). Two of the peridotites contain sparse phlogopite which has given them low ratios (298, 241). Thus two of the rocks have K/Rb comparable with the range suggested by Murthy and Stueber (1967) for alpine peridotites (200-420) and two have much higher ratios, comparable with those

found in mylonitised hornblende-peridotites from St. Paul's Rocks (Hart, 1964). As hornblende may have high K/Rb ratios in ultramafic rocks (Hart, 1964) and other common minerals in ultramafic rocks, olivine, pyroxenes and garnets have low ratios (Allsopp et al., 1969), the high whole rock K/Rb ratios are dependant on the presence of hornblende. An analysed hornblende from specimen 5552B has a K/Rb ratio of 900 and has a ferroan pargasite composition (see Table 2).

In the absence of hornblende from the ultramafic rocks, K/Rb, Al_2O_3 and alkali content would be lower, more comparable with normal types, and limited by the stability of pyroxenes. Hart (1964) found that in the St. Paul's Rocks hornblende-peridotites, K/Rb increased with K, contrary to the trend in oceanic tholeiites. Based on the fact that many amphiboles have K/Rb ratios similar to ocean tholeiites (i.e. in the region of 1500) Hart et al. (1965) suggested that variations seen in oceanic tholeiites could be due to variations in the amount and composition of amphiboles in a parental hornblende-peridotite upper mantle. A similar hypothesis has been forwarded by Oxburgh (1964).

Petrographic evidence suggests that orthopyroxenes in the Coll and Tiree rocks were highly aluminous and calcic, now containing exsolved spinel and pale hornblende, indicative of their removal from a very high P-T regime to a more moderate one, probably on emplacement into the complex. The paragasitic hornblende, whilst in equilibrated polygonal aggregates, shows disequilibrium textures to orthopyroxenes and olivines. Such relationships suggest that the hornblende-peridotites have partially re-crystallised, a matrix of hornblende containing relics of the earlier primary minerals. As olivine and orthopyroxene are still present in the rocks, without evidence of reaction relationships between them, examination of the binary system forsterite-enstatite shows that such a co-existence is only possible under high temperatures and pressures (see Tarney, 1964). It follows that the rocks must have been emplaced under high grade metamorphic conditions, in a similar situation to that found with Scourie picrite dykes in Assynt (Tarney, 1964). Whether this emplacement was prior to granulite facies or during granulite facies, if the ultramafic rocks had had normal original chemistries, then ionic exchange with country rock must have given them their peculiar element abundances, caused or facilitated by the growth of hornblende on recrystallisation. In the granulite facies such an enrichment in alkalies and related elements is

at variance to the observed features in other rocks, which seem to have been purged of these elements, as is the growth of a hydrous phase. Two equally likely cases remain to explain the present chemistry of the fresh hornblende-peridotites. They may have remained a closed system, retaining original chemistry, fluids stable at depth causing hydroxylation of pyroxenes on emplacement. Hornblende may have grown in the rocks in response to influx of H_2O and other elements either before the granulite facies, or because of their composition, the rocks may have taken up migrating material during the granulite facies metamorphism.

The green hornblende-orthopyroxene-plagioclase rim to the hornblende-peridotite at Bousd (5552C) has higher SiO_2, Al_2O_3, CaO, Na_2O and K_2O, and lower MgO than the fresh rocks. Its content of Cr and Ni, although lower than in the fresh material is much higher than in metabasites (Fig. 22c) and shows its origin from ultramafic materials. Its high K/Rb ratio largely reflects the absence of biotite and sparse felspar. Its contrasted chemistry must have involved much element re-distribution.

The high Cr and Ni contents of the hornblende- and biotite-rich outer margins (5552D and E) shows they have been derived from the ultramafic mass. The prominence of biotite in them has imparted low K/Rb ratios and high Rb/Sr ratios (Figs. 24b, 25e). The alteration at amphibolite facies must have involved the influx of H_2O and alkalies, and the loss of Mg and Fe to give the present compositions.

The main points of interest arising from this section are:

1) In all cases, granulite facies representatives of the groups have low K_2O and Rb, but high K/Rb ratios. Ba/Rb is also high and K/Ba and Rb/Sr are low compared with amphibolite facies rocks.

2) Some degree of control has been exerted by major element chemistry on metamorphic mineralogy.

3) Trace element abundances and trends in metabasites are consistent with an igneous origin, although equally may reflect metamorphic variation in mineralogy.

4) Although granofelses show features which might be expected from their close spatial association with metabasites and derivation by some igneous process, they show some marked divergences from this hypothesis. They bear only a passing chemical resemblance to garnet-biotite-gneisses.

Although somewhat basic and alkali-rich, granofelses compare fairly well with average analyses of orogenic zone andesites and dacites.

5) Fresh ultramafic rocks are closely comparable with other Lewisian examples, but show some marked differences from normal alpine-type peridotites. Although their mineralogy and to some extent their chemistry is akin to that suggested for upper mantle materials, it is probably largely a reflection of their metamorphosed state.

6) The abundances of such elements as Rb, Sr, Y, Ce and La are largely explicable in terms of their stability in metamorphic assemblages, and metamorphism has probably obscured original features.

Granitic rocks.

Three groups of quartzo-felspathic rocks which have a granitic appearance are geochemically described. They are distinct in time of origin, and in their chemical composition. Description of concordant, pre-granulite facies granitic veins from granofelses, metabasites and garnet-biotite-gneisses is followed by brief accounts of the geochemistry of discordant quartzose pegmatites and biotite-pegmatites.

Concordant granitic veins.

The mean composition of 21 analysed veins is compared in Table 10 with that for 8 leucogranites from the Snowy Mountains of Australia (Kolbe and Taylor, 1966), the other granitic rocks of Coll and Tiree and Taylor's (1965) estimate of trace element abundances in granites.

It was suggested in Chapter 3 that from field and petrographic evidence, an injection or metasomatic origin for these granites was unlikely, as was an origin by metamorphic differentiation.

The average bulk composition of the veins is very acid, and apart from relatively high CaO content is similar to that for leucogranites from the Snowy Mountains. However due to the large variations in proportions of microcline, depending on the deformed state of these rocks, the group shows great variations in the proportions $CaO : Na_2O : K_2O$. This is well shown by An-Ab-Or plots (Fig. 13c) where undeformed examples cluster close to the Ab-An join at an Ab/An ratio of about 2.0, and a regular trend towards the Or apex is maintained by progressively more deformed specimens. The

TABLE 10 : Average analyses of granitic rocks

	1	2	3	4	5
SiO2	73.8	75.75	85.7	71.8	
TiO2	0.08	0.09	0.02	0.26	
Al2O3	15.5	13.0	8.3	15.3	
Fe2O3	0.57	1.18	0.22	0.85	
FeO	0.63		0.38	0.76	
MnO	0.02	0.03	0.00	0.01	
MgO	0.3	0.2	0.2	0.3	
CaO	2.27	0.71	0.26	0.90	
Na2O	3.9	3.6	4.4	3.5	
K2O	3.24	4.66	1.15	5.83	
P2O5	0.032	-	-	0.037	
Total			100.6		
Trace elements in parts per million					
Cr	16	4	8	5	4
Ni	15	2	2	3	0.5
Cu	116	2	-	9	10
Zn	10	-	-	9	40
Rb	88	388	35	130	150
Sr	247	42	100	361	285
Y	8	26	1	5	40
Zr	106	88	32	487	180
Nb	4	-	1	4	20
Ba	247	270	427	1985	600
La	11	30	1	245	25
Ce	23	-	1	359	46
Pb	46	23	14	19	20
Th	8	17.2	1	33	17
K/Rb	306	100	272	372	-
Ba/Rb	5.15	0.70	12.2	15.26	4.0
K/Ba	108.8	143.3	22.4	24.38	-
Rb/Sr	0.356	9.2	0.35	0.360	0.526
Ba/Sr	1.00	6.43	4.27	5.50	2.11
Ca/Y	2025	195	-	1285	-
Ce/La	2.09	-	-	1.47	1.84

1. Average analysis of 26 concordant granite veins : Coll and Tiree.
2. Average analysis of 8 leucogranites: Snowy Mountains (Kolbe and Taylor, 196).
3. Analysis of quartzose pegmatite 4932B : Coll.
4. Average analysis of 5 biotite-pegmatites : Coll.
5. Estimated trace element abundances in granites (Taylor, 1965).

undeformed granites plot remote from the low temperature trough of the system and the trend of increasing K_2O crosses it.

A similar trend of increasing K_2O is displayed by a Q-Ab-Or diagram (Fig. 23a). On this, the undeformed specimens cluster close to the Q-Ab side, and the trend of increasing Or content proceeds towards the Or apex, crossing the granite maximum region. A garnet-orthopyroxene-granite (78SE1C) from Hynish is considerably affected by its content of mafic minerals, and plots towards the Ab apex. The cluster of undeformed samples is consistent with a cotectic line at 1 kbar vapour pressure and an Ab/An ratio of 2.0 (Winkler 1967), it is less potassic than the ternary minimum of this curve. Such a situation could arise by extensive melting of a K_2O-poor host rock, the melting trend beginning at the ternary minimum and following the cotectic line in the system. At that point where the tie between host rock and an apex of the system crossed the cotectic, the composition of the melt would proceed along the tie-line until it reached the plot of the host rock's original composition at the point of complete melting. Combined plots of spatially related pairs of host rock and granite vein on Q-Ab-Or show no meaningful relationships. Both rock types suffered element re-distribution during subsequent metamorphisms, and any characteristic chemical relationships have been obscured.

Further evidence of the variation of K_2O in the granitic veins is shown by plots of Niggli k against mg (Fig. 16c).

On an Fe-Mg-Alk diagram (Fig.14c) the granites plot close to the Alk apex as expected, and maintain a fairly constant Fe/Mg ratio of about 4.0, a result of the content of garnet in many of the specimens.

The low abundances of Cr and Ni (16 and 15 ppm) are not incompatible with the granitic veins being high Ca-granites or granodiorites for which Taylor (1965) quotes average values of 20 and 20 ppm. They are however too low to be consistent with a metasomatic or metamorphic segregation origin as the associated rocks have contents an order of magnitude higher and would influence any metasomatic development in them.

Rb shows wide variation in these rocks (Fig. 18c) from 4-335 ppm, positively correlating with K_2O and negatively with Na_2O and CaO. This indicates that it increases with microcline at the expense of plagioclase. Its low abundance in most of the rocks, which also have high K/Rb ratios, is at variance with data from highly differentiated granites like the Snowy

Mountains leucogranites (Table 10) which are enriched in Rb relative to K. This would rule out the veins' origin by influx of differentiated melts or permeating granitic fluids as White (1966) has deduced from similar veins in Australian migmatites.

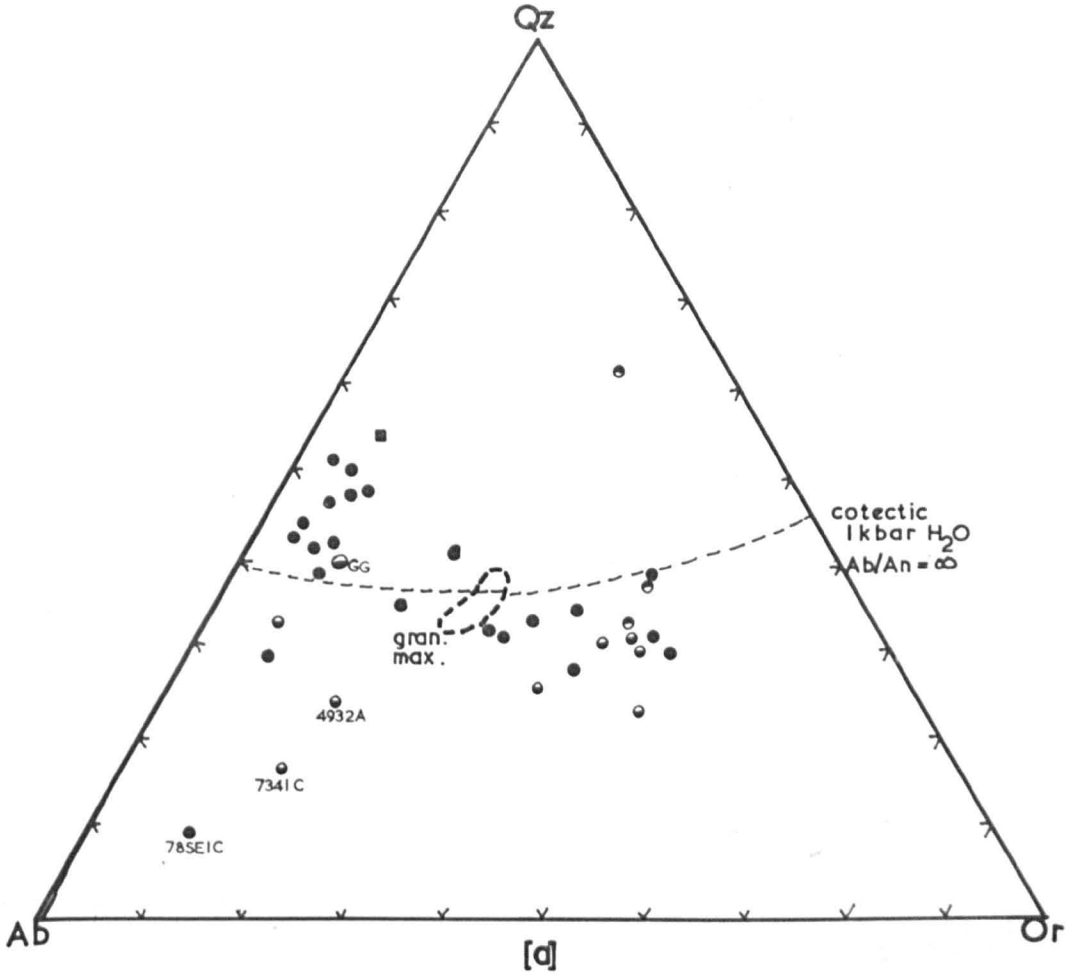
Had the granites formed anatectically before granulite facies metamorphism, leaving a biotitic residue, they would have had an initially higher K/Rb ratio than their parent rocks. On the other hand, if the residue was dominantly hornblende-bearing, the granites would have had initially low K/Rb ratios, the residue acquiring a progressively higher ratio dependant on the amount of melt formed. K/Rb ratios equal to those of their host rock could only have developed had an alkali-poor residue of a pyroxenic nature been left after partial melting at the onset of, or during granulite facies metamorphism. A plot of K/Rb for granites against K/Rb for adjacent host rocks (Fig. 23c) shows that in general, the veins have the higher ratios. This is to be expected where the veins' felspars can have high K/Rb ratios compared to biotites in granofelsic host rocks. Where amphibolite is the host rock, the opposite is the case. The relationships shown are probably due to equilibration of Rb between microcline in granite and biotite or hornblende in host rock during the amphibolite facies.

In general the least deformed granites show the highest K/Rb ratios as well as low K_2O (Fig. 24c). They also have high Ba/Rb and low Rb/Sr ratios.

Both Sr and Ba are very variable in the granites, Sr following Ca (Fig. 19f) and Ba following K (Fig. 19c). As a result all ratios involving K, Rb, Ba and Sr have high variation.

Some interesting features of the granites are their abnormally high abundances of Pb and Cu (46 and 116 ppm) and the variability of Th from 0 to 39 ppm. In view of the fact that lead mines were once worked at Crossapol these high values might indicate selective mineralisation of the granitic veins at some stage. Neither Pb nor Cu show any significant correlations.

Y is very variable (0-34 ppm) and highest in garnetiferous varieties, being concentrated in garnets (Table 2). La and Ce are variable too, (0-67 ppm and 0-102 ppm) generally being highest in deformed and remobilised concordant granites. Unlike in the biotite-pegmatites (see below) the Ce/La ratios are normal at about 2.0 (Fig. 23b).



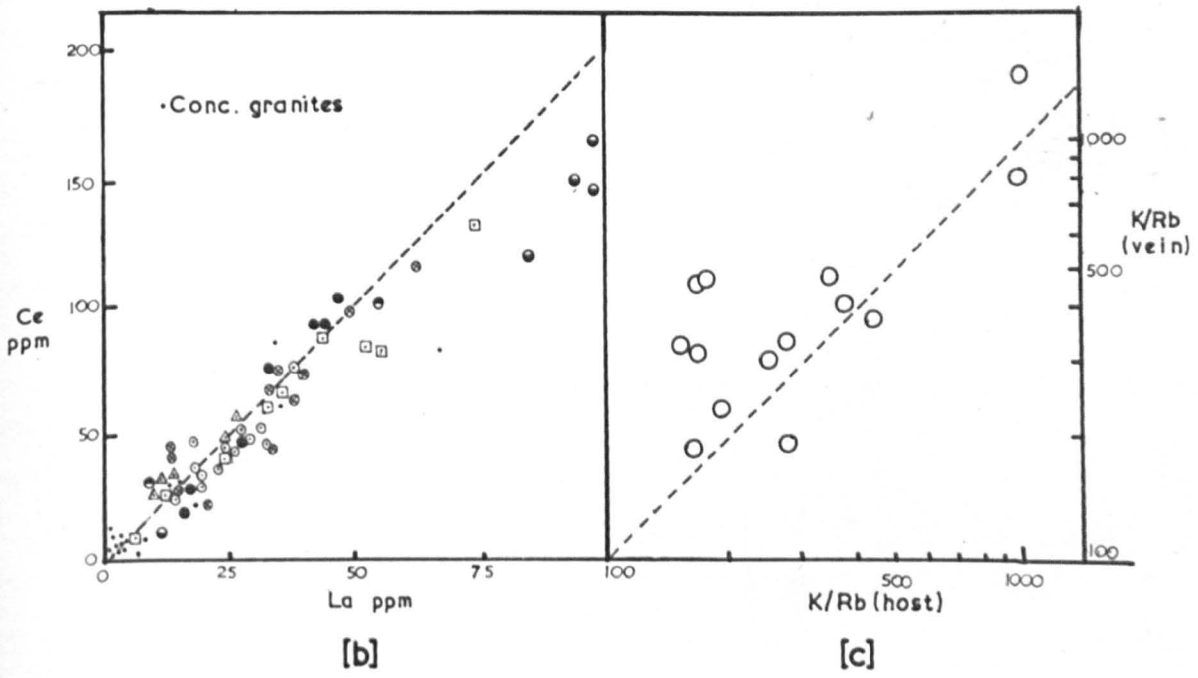


Fig.23: Q - Ab - Or diagram for granitic rocks, graph of Ce against La for gneisses and granitic rocks, and graph of log K/Rb for concordant granites against log K/Rb for their host rocks. Symbols as Fig.13 mainly.

Quartzose pegmatites.

One analysis of an undeformed quartzose pegmatite is shown in Table 10. Its quartzose nature is indicated by its very high SiO_2 content. On a Q-Ab-Or diagram (Fig. 23a) it plots close to the Q-Ab side with a high proportion of normative quartz. Its very CaO-poor nature is shown on an An-Ab-Or plot (Fig. 13c) where it plots on the Ab-Or side close to the Ab apex. Compared with the other granitic rocks and normal granites, the quartzose pegmatite has a high Na/K molar ratio of 85. This, coupled with its very high quartz content, enables its origin to be discussed.

Bowes (1967) held that autochthonous quartz-plagioclase veins in the Mainland Lewisian were products of supercritical solution in the amphibolite facies. From experiments on the solubility of albite in supercritical hydrous fluids, Currie (1968) found that the ratio SiO_2/Na_2O in the solution varied with temperature and pressure. Between 400 and 600°C, the ratio varied from 4.6 to 19, SiO_2/Na_2O in albite being 7.8. Na_2O was present in excess of albite composition below 550°C but became rapidly less as temperature increased. He concluded that veins produced by supercritical solution in metamorphic rocks would be albite-rich in low grade rocks, but quartz-rich in high grade. The relative age of the quartzose pegmatites, between D_2 and D_3, is consistent with conditions changing from anhydrous to hydrous during D_3.

Biotite-pegmatites.

The mean of 6 analyses of discordant biotite-pegmatites and granites is given in Table 10. They have been plotted with other granitic rocks on various diagrams.

In Chapter 3 evidence was presented, indicating that these rocks are products of crystallisation of fluids, having a coarse polygonal texture. Those specimens which have been analysed are largely unaltered. Apart from 64NE1A they cluster well on the triangular diagrams of An-Ab-Or, Fe-Mg-Alk and Q-Ab-Or (Figs. 13c, 14c, 23a). Also shown on these plots, and others are discordant pegmatites from metasediments (7341C) an epidotic crush zone (4932A) and an agmatite (GG). They are not described in the text.

In major composition the biotite-pegmatites compare well with leucogranites from the Snowy Mountains, being perhaps a little more basic and certainly more potassic. They are easily distinguished from the concordant granites by their much lower CaO contents (Fig. 19f).

Five of the biotite-pegmatites show on Q-Ab-Or and An-Ab-Or diagrams as a cluster close to the potassic end of the concordant granites' trend and approximate the composition of deformed and remobilised concordant granites. They are too rich in K_2O to fall near to the thermal trough on An-Ab-Or or the granite maximum on Q-Ab-Or, and as they are undoubtedly crystallised products of a melt or fluid must be regarded as K_2O-enriched, relative to normal granites. In this respect they resemble intrusive Lewisian granites from the Mainland (Bowes, 1967).

Their contents of Cr, Ni, Cu, Zn, Rb, Sr, and Pb are normal and not particularly variable.

There are several points of interest arising from their trace element geochemistry. Their K/Rb ratios are high (Fig. 18c), approximately those of the concordant granitic veins, and are not consistent with their origin by a normal fractionation process.

Sr contents are the same as in the concordant granites although there is no correlation between CaO and Sr. Most of the biotite-pegmatites have low CaO/Sr ratios and compare closely with remobilised concordant granites in this respect (Fig. 19f).

The biotite-pegmatites have high Ba content and also high Ba/Sr ratios, a feature usually associated with late stage fractionates, the larger size of Ba favouring its retention in differentiating magmas. However, their Rb/Sr ratios are not excessively high (Fig. 25f) as would be expected from a Rb enriched, and Sr depleted late fraction of Shaw's (1968) pegmatite-hydrothermal trend. K/Ba ratios are normal and similar to many from the concordant granites although not as high as the deformed varieties (Fig. 24f). Presumably Ba is present in the dominant alkali felspar, Sr being present in smaller quantities in this and in subordinate plagioclase (Sr is almost absent from micas).

Although in most examples Zr abundances are normal, in an allanite-bearing example it is as high as 2093 ppm, an enrichment of a factor often compared to normal values around 200 ppm. Abnormally high Zr is usually restricted to very late stage differentiated granites (Taylor, 1965).

In all the examples the rare earth elements Ce and La are very abundant, especially in the allanite-bearing specimen 5573B where they are at levels of 1679 and 1168 ppm respectively. Now under conditions of extreme fractionation, the small differences between ions of La and Ce are emphasised, and the larger La ion should be enriched relative to Ce. In all the specimens of biotite-pegmatite the Ce/La ratio is low relative to normal crustal values of about 2.0. The biotite-pegmatites are compared with concordant granites and gneisses on a graph of Ce against La (Fig. 23b).

Th is concentrated to a level of 145 ppm in the allanite-bearing specimen being contained in allanite at extremely high concentrations (Table 2). Again this is a feature of late stage hydrothermal or pegmatitic fractionation.

Noteable points arising from thin section are:

1) The extreme variability with respect to certain chemical parameters of the concordant granitic veins, many being functions of metamorphic grade and state of deformation.

2) The incompatibility of the chemistry of the concordant granites to an origin by metasomatism, intrusion or metamorphic differentiation.

3) The SiO_2- and Na_2O-rich nature of the quartzose pegmatites.

4) The persuasive similarities between biotite-pegmatites and deformed and remobilised early concordant granites.

5) The highly fractionated nature of the biotite-pegmatites with respect to concentrations of Ba, Zr, Ce, La, Ca/Sr, Ba/Sr and Ce/La ratios.

Mylonitic rocks.

An analysis of a black flinty crush rock in migmatitic gneiss (Table 11) compares reasonably with mean analyses for the migmatites, but is more basic. Relative to the migmatitic gneisses it has higher total Fe and MgO, and lower SiO_2, Al_2O_3 , Na_2O and K_2O. Its trace element content shows no marked differences, but is richer in Cr and Ni, poorer in Rb and Ba, and has an anomalous low Ce/La ratio of 0.6. Its basic composition might reflect the loss of granitophile elements during its formation by frictional fusion. It has comparable K/Rb to the migmatitic gneisses. In Table 11 it is compared with pseudotachylite from Meall Deise (Bhattacharjee, 1963).

TABLE 11 : Analyses of mylonitic rocks

	1	2	3
SiO2	61.41	49.55	53.8
TiO2	0.60	0.28	1.1
Al2O3	14.9	15.4	16.8
Fe2O3	5.25	10.70	3.2
FeO	2.61	0.73	5.5
MnO	0.68	0.00	0.12
MgO	3.7	0.3	1.9
CaO	3.87	21.25	8.2
Na2O	2.9	0.0	4.5
K2O	1.75	0.46	3.4
P2O5	0.103	0.072	0.58
H2O	2.00	1.93	0.81
CO2	1.00	1.10	0.40
Total	100.09	101.75	100.3

Trace elements in parts per million

Cr	231	28
Ni	109	11
Rb	56	9
Sr	338	979
Y	10	8
Zr	136	57
Nb	3	3
Ba	492	116
La	24	10
Ce	38	28
Pb	19	12
Th	1	4

1. Black flinty crush rock 4024 : Coll.
2. Green epidotic crush rock 5577 : Coll.
3. Pseudotachylite : Meall Deise, Gairloch (Bhattacharjee, 1963).

An epidotic crush rock analysis reflects its high epidote content, being depleted in alkalies and MgO and very rich in total Fe, CaO and Al_2O_3. Its trace element content is noteable for very high Sr relative to Ba. Sr probably substitutes for Ca in epidote.

Origins.

Peach et al. (1907) referred to the bulk of the Scottish Lewisian as "orthogneiss". Few authors now are satisfied with such an unspecific term, and have tried to explain its origin, and that of similar gneiss terrains by several hypotheses, to a greater or lesser extent hinging on geochemical evidence.

Many consider the possibility of a large sedimentary component (e.g. Sutton and Watson, 1951a; Dearnley and Dunning, 1968). To some extent this view is supported by transitions from recognisable sediments into quartzo-felspathic gneisses, as in South Harris (Skinner, 1970), Gairloch (Park, 1963) and the Cascades area (Misch, 1968). The presence of great thicknesses of Precambrian grey wackes of suitable composition in Wyoming (Condie, 1967) encourages speculation as to the results of high grade metamorphism on such rocks. Undoubtedly, if the Lewisian gneisses and other Shield gneisses are metasediments, then they can only represent vast accumulation in geosynclinal environments, as uniform grey wackes could only form the present uniform gneisses. Condie's (1967) mean for Wyoming grey wackes is given in Table 5 for comparison with the Coll and Tiree gneisses in Table 3.

Hypotheses for gneisses' origins from supracrustal volcanogenic sequences have been encouraged by Goodwin's (1968) description of large volumes of ancient volcanic rocks of a calc-alkaline suite from the Canadian Shield. He suggested an early crust forming process involving the building up of a "platform" of predominant tholeiitic basalts, followed by an "edifice stage" of accumulation of mainly pyroclastic rocks of calc-alkaline nature, becoming more felsic with time. The volcanic pile thus erected was subjected to erosion giving thick volcanogenic sediments. Bowes (1968a) recognised the possibility that large proportions of extrusive igneous rocks could be present in the Scourian, Kylesku Group. The abnormally low K, Rb, Rb/Sr and initial Sr^{87}/Sr^{86} ratios (Evans, 1965)

and very high K/Rb ratios of Assynt gneisses have encouraged Sheraton (1970b) and Skinner (1970) to postulate their origin from an andesitic pile. The low initial Sr^{87}/Sr^{86} ratios of these Scourian, granulite facies gneisses could only allow normal Rb/Sr ratios, of about 0.20 to be present for a short time prior to Rb's being purged from the system during the granulite facies. They sought a parental material having low K, Rb and Rb/Sr ratios. Condie's (1967) Wyoming grey wackes, largely unmetamorphosed, still have normal Rb/Sr ratios and were rejected. A convenient source was found to be andesitic lavas. Taylor and White (1965) and Taylor et al. (1969) described andesites from the island areas of the West Pacific, having low K, Rb, Rb/Sr ratios averaging 0.07, and high mean K/Rb (430). Both Sheraton and Skinner suggested that many of the chemical features of the Drumbeg and Assynt gneisses could have been derived by metamorphism of such andesites and their loss of K and Rb at granulite facies. Furthermore, the very low Rb/Sr ratios of andesites would not necessitate Rb loss from the system at granulite facies for some considerable period (1000 myr) after the original formation of the volcanic pile. This, on their account, would be a pre-requisite for the retention of low initial Sr^{87}/Sr^{86} ratios.

Holland and Lambert (1969), considering the high K/Rb and low Rb/Sr and K/Pb ratios of Scourian gneisses, suggested that these rocks might have differentiated directly from the mantle, retaining the extreme values of the above parameters thought to exist in the mantle. They further suggested that the present layering and shallow dips of Scourian rocks could have been formed by strain tangential to the mantle/crust boundary in a low viscosity layer, or one of uniform rheology, characterising the high pressure granulite facies.

In view of the fact that on Coll and Tiree, both gneisses and large belts of undoubted metasediments were in place before basic dyke intrusion and thus before granulite facies metamorphism, and the polyorogenic nature of their history, it does not seem realistic to draw inferences from the gneisses' geochemistry as to their ultimate origins. In many respects the Tiree hypersthene-gneisses are similar to the Scourian gneisses of Assynt, but are much more potassic. Skinner (1970) has drawn a distinction between the Scourian gneisses of Assynt and rocks of similar age on South Uist and at the Butt of Lewis in terms of K/Rb ratios. Whilst the high

mean K/Rb of Assynt gneisses may indicate their origin from andesites, the lower values of the last two areas possibly reflect an origin from grey wackes with normal lower initial K/Rb ratios. He has neglected to indicate that Scourian metamorphic conditions may have varied in the Lewisian, with consequent effects on the intensity of element migration.

Whilst a sedimentary origin for the thick belts of quartzites, marbles and calc-silicate rocks on Coll and Tiree is most probable, those rocks which are generally found as thin bands in the dominant gneisses, such as garnet-biotite-gneisses, may have different origins. It does not really seem valid to assign a sedimentary origin to a metamorphic rock either on the basis of a strange chemical composition or the presence of such minerals as garnet, Al_2SiO_5 polymorphs or anthophyllite, unless a metamorphic progression can be established from unaltered sediments to metamorphic rocks. The contrast of, say, garnet-biotite-gneiss to quartzo-felspathic gneiss may have arisen through non-sedimentary processes in the early history of the complex. A parallel dilemma is found in the case of the Precambrian Dutch Gin Schists of Pembrokeshire which, being garnetiferous, were long regarded as sediments. However, re-examination by Baker et al. (1968) has shown that they properly represent recrystallised mylonites, derived along a thrust cutting a diorite intrusion. Thorpe (1970) has shown the strong chemical dissimilarities between the two genetically linked rock types. Furthermore Sutton and Watson (1959) have described recrystallised mylonites from deep seated transcurrent fault zones in charnockitic rocks in Tanzania, which contain garnet and kyanite, and are presumably aluminous. In the author's own experience garnet growth in Caledonian mylonites from the Moine Thrust Zone is quite common. Considering the undoubted progression of the Lewisian of Coll and Tiree, through lower grades before granulite facies metamorphism, any of a large number of processes could have produced chemical anomalies in otherwise monotonous lithologies by redistribution of Fe, Al and Si in particular. If the multi-phase large scale deformation suffered is taken into account, such inhomogeneities could have had any original shape and then have been deformed to planar bodies.

As regards the metabasites, their chemistry, particularly their contents and variations in the elements Mg, Ti, P, Ni and Cr, is compatible with their origin as intrusive, possibly differentiated basic igneous rocks. The ultramafic masses with their high Cr and Ni contents, peridotitic comp-

osition, and frequent association with probably incompetant metasediment bands, and in the case of the Loch a Phuill mass with garnet-biotite-gneiss, either incompetent metapelite or recrystallised mylonite, suggests their origin as Alpine-type ultramafics; i.e. tectonically emplaced. The metamorphosed nature of both groups has masked their original chemical characteristics. Presumably, if the literature is to be believed, their ultimate source was the upper mantle. In the case of the ultramafic bodies, their peculiar high abundances of K, Na, Rb, Sr, Ba and Pb, are due to their content of hornblende. Although the hornblende could have formed on introduction of ultramafic material into a higher level than its origin, if it represents the effects of earlier hydrous fluid stable with anhydrous assemblages in the materials at depth, then the present composition may be original and representative of mantle derived material. Tarney (personal communication) has reflected upon the partial melting of an ultramafic mantle material containing hornblende, and its consequent element contents, perhaps giving rise to observed trace element distributions and trends in a suite of Lewisian picrite dykes from Assynt. The intrusive nature of the mafic granofelses and their chemical affinities suggests they be grouped with the metabasites, although they are more noritic.

Intermediate granofelses present a thorny problems, as they show chemical affinities to both metabasites and garnet-biotite-gneisses, and in other respects stand alone. They show a passing similarity to orogenic andesites, but similar arguments to those applied above suggests that their present compositions, especially trace element abundances, are largely due to metamorphism. Their present lack of fabric is of fundamental importance in assigning an origin to the granofelses, and their parental material must have been similarly homogeneous. The granofelses could represent uniform semipelitic sediments, andesitic or dioritic intrusions, or aphanitic mylonites formed early in the history of the complex and subsequently recrystallised. It is worth noting that flinty crush rock from Coll is more basic than its parental gneiss, as are pseudotachylites from Meall Deise (Bhattacharjee, 1963), and both show some chemical similarity to granofels (Tables 8 and 11). The high Cr and Ni content of granofelses compared with quartzo-felspathic gneisses are also present in flinty crush rock derived from the gneisses.

Field, petrographic and to a limited extent, geochemical evidence from concordant granitic veins in granofelses, metabasites and garnet-biotite-

gneisses, indicates their incompatibility with a formation by metasomatism, intrusion or metamorphic differentiation. As with the other pre-granulite facies materials, their composition is probably now different from the original, but the weight of evidence indicates that they formed either by selective melting or solution from their host rocks.

The origin of the biotite-pegmatites is dealt with in discussion of geochemical changes associated with amphibolite facies retrogression.

Geochemical Changes at Granulite Facies.

Representatives of the early rock types in the complex, still retaining granulite facies assemblages, all show several chemical characteristics in common which distinguish them from members of their groups at lower grade. Granulite facies rocks have high mean K/Rb and Ba/Rb ratios and low mean K/Ba and Rb/Sr ratios than lower grade rocks. Significant differences in each group are:

i) Hypersthene-gneisses, although having similar K_2O contents to the lower grade rocks, have low Rb, Y and Th, and high Ba and Sr.

ii) Pyribolites and mafic granofelses have low K_2O and Rb. The latter have abnormally high Ba and Sr.

iii) Fresh ultramafic rocks have low K_2O and Rb.

iv) The least deformed and highest grade concordant granite veins have low K_2O, Rb and Ce and La.

These features are displayed by plots of K against Rb, Ca against Sr, and K against Ba (Figs. 18 and 19), and also by K/Rb against K, Ba/Rb against Rb, Rb/Sr against Rb and K/Ba against K (Figs. 24 and 25).

Compared to accepted averages for the whole crust and crustal rocks of comparable compositions the K/Rb ratios of all the granulite facies rocks are abnormally high. Similar abnormal values have been recorded from other areas of granulite facies rocks by Sighinolfi (1968) from the Alps, Sheraton (1970a,b) from Drumbeg, and Skinner (1970) from Assynt. Heier (1960), Lambert and Heier (1968) and Whitney (1969) have also demonstrated depletion of Rb relative to K in granulite facies rocks, and relatively high K/Rb ratios seem to be a feature of such materials.

Taylor (1965) suggested that values of K/Rb acceptable as normal lie between 150 and 300, the mean value for igneous rocks being 230. Values outside this range would call for special explanations. Most anomalies

described at the time of publication were low K/Rb ratios caused by extreme igneous fractionation exploiting the larger Rb^+ ion and concentrating the element relative to K in late fractions. This conclusion was substantiated by Shaw (1968) using covariance analysis on available K and Rb analyses from igneous and quasi-igneous rocks. He showed that in his Main Trend, excluding oceanic tholeiites, for low levels of K of about 0.01 per cent K/Rb rose to 430, although lying within Taylor's limits, for K levels greater than 1 per cent. Using similar techniques, Sighinolfi (1969) showed that in Alpine granulite facies rocks, relative depletion in Rb at low levels of K is even greater. Both Shaw and Sighinolfi then, found that at low levels of K, K/Rb is negatively correlated with K.

Three possibilities exist for the production of high K/Rb ratios in the granulite facies rocks of Coll and Tiree:

a) The rocks initially had high K/Rb ratios.

b) It is a mineralogical feature.

c) Granulite facies metamorphism in some fashion resulted in Rb being depleted relative to K.

As has been mentioned above, hypersthene-gneisses have higher K/Rb ratios than greywackes. The ratios are also higher than those reported from andesitic rocks by Taylor and White (1966) whose mean value is 430. Although basic igneous rocks may have high K/Rb ratios (Gast, 1968), those with high ratios are from oceanic regions. Oceanic tholeiites with their very low K contents (600 ppm) have K/Rb ratios in the range 500 to 2000, but their typical Ba/Sr ratios of about 0.1 rules out any similarity with pyribolites and mafic granofelses whose Ba/Sr ratios are 1.0 and 1.3 (Tables 6 and 8). Furthermore, the intrusion of these rocks into what must have been an orogenic belt rules out the possibility of their being abyssal tholeiites. Continental tholeiites have K/Rb ratios typically below 500 (Murthy and Stueber, 1967). Ultramafic alpine-type intrusions have low ratios compared with the highest recorded from Coll and Tiree.

There remains the possibility that the high K/Rb ratios were inherited direct from the mantle by a process of differentiation. Notwithstanding their close association with supracrustal rocks, the derivation of the hypersthene-gneisses from the mantle would involve differentiation of crustal material enriched in Al, Si, Na and Ca, and it seems improbable that it would not also be enriched in K, Rb, Th and U. Moorbath et al. (1969) have shown that U depletion occurred in the Lewisian about 2900 myr ago, this age being equated with the Scourian orogeny. These authors

favoured rejuvenation of crustal material, rather than upper mantle differentiation, as the most probable mechanism by which the Scourian complex formed. Furthermore, the probability of a metamorphic history before granulite facies metamorphism, and after the laying down of sediments on a mantle-derived basement (which must surely have been uplifted to allow sedimentation) would act on a "protocrust" in a similar fashion to the effect retrogressive metamorphism has had on the granulite facies rocks; i.e. give it normal crustal element abundances. In view of the statements made above about the uncertainty of the origin of the quartzofelspathic gneisses, their formation as a primaeval crust from the mantle at a much earlier stage in the Earth's history cannot be ruled out entirely. However, in view of the necessary uplift prior to the deposition of the undoubtedly sedimentary quartzites, marbles, calc-silicate rocks etc., and then a progression back to the deep crustal level of intermediate to high pressure granulite facies, it is hardly likely that originally mantle-derived material would have any chemical, structural or metamorphic resemblance to its original state.

In considering the possibility that high K/Rb ratios have been produced in the granulite facies rocks from lower grade rocks with more normal K/Rb ratios by redistribution alone of Rb between coexisting K-rich minerals, those rocks crystallising with minerals of high K/Rb must be balanced by rocks in which Rb is enriched relative to K in minerals if no Rb has been lost to the system as a whole. The analyses do not bear this out, no granulite facies rocks of any type were found with abnormally low K/Rb ratios. Amongst common metamorphic K-rich minerals only micas have the large K sites (of 12-fold co-ordination) favouring Rb enrichment. Micas are generally thought to be unstable in the granulite facies, and certainly they are rarely present in the rocks under consideration.

There remains the case of strong depletion of Rb relative to K, Th, K and perhaps Y in the granulite facies. In most of the documented cases of granulite facies rocks with high K/Rb, the various authors have favoured removal of Rb from the system by some process, and suggest that such processes are typical of conditions in intermediate to high pressure granulite facies terrains.

Lambert and Heier (1968) concluded that in the medium to high pressure granulite facies rocks of SW Australia there was a trend towards lower abundances of Th, U, Rb, Pb, K and Si and higher Ca, Mg, Fe, Mr, Ti etc.

compared with the more common amphibolite and low pressure granulite facies shield rocks. They suggested that this trend was typical of deep crustal material. It should be noted that in their case, the abundances of Rb, K, and Th, and ratios of K/Rb, Rb/Sr etc. (Table 4) did not have anywhere near the extreme values as is the case in these Lewisian rocks. Mean K/Rb for instance only varied from 200 to 309 from amphibolite to granulite facies rocks in the Musgrave Range, and if acid gneisses alone are considered no significant difference between facies can be seen. However, to explain what element fractionation they had observed they suggested the main process to be one of anatexis of the gneisses, this being supported in the main by paucity of acid (> 70 per cent SiO_2) rocks in the granulite facies terrains which might be a result of removal of melt phase. They provided the alternative hypothesis that sub-solidus transport of granitophile elements became very efficient in the medium to high pressure granulites.

Briefly, Whitney (1969) in his consideration of K/Rb trends in the Adirondack paragneisses suggested that the high ratios in granulite facies rocks were possibly due to Rb concentration in anatectic melts. He pointed out that such anatectic depletion would depend on the stability of biotite at the onset of anatexis. If biotite remained in the residue it would concentrate Rb and the melt phase would have high K/Rb. If biotite broke down then the melt phase would concentrate Rb and leave a residue with high K/Rb. White (1966) used similar reasoning in considering high K/Rb of granitic veins in biotite-rich migmatites of the Palmer Region of South Australia.

Sighinolfi (1969) found that there was a difference in mean K content and K/Rb ratios between amphibolite and granulite facies rocks in the Alps (K=2.65 and 1.53 per cent; K/Rb = 231 and 505, respectively). By regression analysis he established that there was a significant negative correlation between K/Rb and K, so that he postulated removal of K and Rb at granulite facies, Rb being preferentially depleted. Again, an anatectic process was seen as responsible.

In accounting for the high K/Rb ratios of Lewisian gneisses from Drumbeg and Assynt, Sheraton (1970a) and Skinner (1970) suggested that they were partly inherited from andesites with moderately high K/Rb ratios. Anatectic processes during the Scourian granulite facies completed the

elevation of K/Rb ratios by selective removal of Rb relative to K.

Now although anatectic removal of alkalies from the lower crust (i.e. medium to high pressure granulite facies) is an attractive hypothesis, leading to the recognised more basic nature of granulite facies gneisses, it does not satisfactorily fit all the evidence. In the Lewisian granulite facies rocks, including those of Coll and Tiree, there is little visible evidence of relics of anatectic melts in the rocks. That which is present; discordant garnet-pegmatites in hypersthene gneisses, and the concordant granitic veins mentioned, neither show any relationship to anatectic rocks on triangular diagrams, nor do they have low K/Rb ratios, rather the opposite. Sheraton (1970) has described deformed microcline-gneisses from the Drumbeg gneisses which he suggested were deformed pegmatites; apart from high K content, their geochemistry is the same as that of country gneisses and they have similarly high K/Rb ratios. Thus what examples there are to suggest anatectic processes operative in the granulite facies themselves have suffered the same depletion as their host rocks. Further evidence against a solely anatectic process lies in the fact that basic and ultrabasic rocks too have higher K/Rb ratios than normal crustal values for their equivalents, and their representatives at amphibolite facies. They too have been depleted in Rb and K, although their K/Rb ratios show a marked trend of decreasing with K (Fig.24b) which is returned to later.

Heier (1965) has neatly expressed the complexities of element movement in metamorphism, and points out the nonexistence of isochemical metamorphism in the presence of a hydrous disperse, or intergranular, phase. Most metamorphic mineral reactions are not simply re-organisations of atoms within silicate frameworks, but involve transitions through the disperse phase. Any mineral transformation in a rock is chemically contributed to by most other minerals in a rock. Metamorphic recrystallisation must result in a number of elements being removed from crystal lattices so that they are easily available to transporting agencies. The elements liberated must enter independant mineral structures, adhere to mineral surfaces or disperse themselves throughout the rock. In dynamic regional metamorphism conditions may exist in which those elements incompatible in stable minerals migrate in or through an intergranular phase, (Mueller, 1967; Gresens, 1966).

A first basic tenet for the observed features, especially with reference to K/Rb ratios, in the granulite facies rocks of Coll and Tiree must be; their trace element compositions are determined by those of stable minerals present. In brief the following minerals present in the rocks contain

sufficiently large amounts of K and Rb to significantly effect the K/Rb ratios of the whole rocks;

Hypersthene-gneisses : untwinned K-felspar, high temperature plagioclase containing up to 9 per cent K-felspar in solid solution according to Sen (1959).

Concordant granites : antiperthitic plagioclase, subordinate untwinned K-felspar.

Pyribolites and mafic granofelses : plagioclase and brown hornblende.

Ultramafic rocks : pargasitic hornblende.

Of those K-minerals possibly in existence before granulite facies, plagioclase and K-felspar would remain stable, hornblende and mica would tend to break down. Their breakdown would essentially release their constituent elements to a disperse phase. Some of these elements would be taken up in rapidly succeeding high grade mineral growth, some would be incompatible in high grade minerals. Considering K, this might be lost to the system or might help form new K-felspar depending on the reactions possible in rocks of particular chemistry. Rb released would have to compete with other elements of similar size for entry into new phases or the pre-existing ones, as it does not form a separate mineral.

Data for ranges of K/Rb given by various authors and some determinations contained in this work are given below for the K-bearing minerals stable at granulite facies.

	K-felspar	Plagioclase	Hornblende
Heier, 1966	560 - 784	780 - 2100	
White, 1966	245 - 554		
Hart et al., 1966		280 - 342	100 - 5000
Phillpotts and Schnetzler, 1970	470	281 - 3610	
This work			
pyribolite 78SW5B		423	2426
ultramafic 5552B			1006

Phillpotts and Schnetzler (1970) made the interesting point that K/Rb ratios in plagioclase phenocrysts, relative to their matrices (melts) increase with increasing alkali content. Thus An_{93} has a K/Rb ratio of 281 and An_{65}, 3610. Their main conclusion was that the distribution coefficient for K/Rb between plagioclase phenocrysts and their matrices is high, averaging 3.0. Metamorphic plagioclases may have been in a similar equilibrium with a disperse phase, and it follows that they may greatly exceed such an intergranular phase in K/Rb ratios.

In hornblendes and plagioclases, both K and Rb must compete for sites which are in general more easily occupied by such elements as Ca and Na, of smaller ionic radii. Because of the relatively smaller size of K^+ ions, they will be strongly preferred, in such substitutions, to Rb^+ hence the high K/Rb ratios of these two minerals. Now, in plagioclase, Rb^+ also has a competitor in Sr^{2+}, which usually substitutes for Ca^{2+}. The smaller ionic radius of Sr^{2+} and its stronger bonding will greatly influence its compatibility in plagioclase relative to Rb^+, and tend to give very low Rb/Sr ratios in that mineral.

K-felspars obviously concentrate K^+ ions, which may be replaced in the lattice by Rb^+, Ba^{2+}, Sr^{2+} and Pb^{2+} to a certain extent, all of which have a radius close to that of K^+. Now, in the presence of large concentrations of Ba and Sr (as in the hypersthene-gneisses), Rb will be at a disadvantage to these elements, in substituting for K^+ and occupying K^+ sites. The smaller ionic radiis and stronger bonding types of Ba and Sr will allow them to enter K sites, and perhaps to cause a certain amount of collapse of the felspar lattice, thereby excluding Rb. In the absence of any data on differences in the lattice between untwinned K-felspar and microcline it is difficult to make any statement on the feasibility of such a process. However, at the supposed deep crustal levels attributed to medium to high pressure granulite facies metamorphism, it is probable that much disorder due to lattice collapse is present in the stable K-felspar lattices. Such a situation would favour concentration of Ba and Sr in K-felspar and place Rb in an incompatible state in that mineral.

It is well accepted (Ramberg 1952, Heier 1965) that in the granulite facies the main effect of high pressures and temperatures is the destruction of hydrous phases such as biotite and hornblende, and the formation of anhydrous assemblages, although because of a progressive change in composition or varying dehydration some hornblende may remain stable. As a result,

the disperse phase becomes more hydrous and under the influence of pressure and temperature gradients will migrate upwards. Some K and Rb would have been added to the disperse phase by breakdown of hornblende and micas. It has been shown that compositional gradients will exist across certain stable mineral grain boundaries tending to cause Rb to move into the disperse phase. It will be accompanied to a certain extent by K, depending on the reactions during the transformation from hydrous to anhydrous assemblages. A tendency for Ba and Sr to be accepted into K-bearing minerals will be developed, or at least these minerals will not lose Ba and Sr. The nett result then is a migrating disperse phase enriched in K and Rb especially. Although a dynamic equilibrium with respect to cations will exist between minerals and disperse phase, more Rb will leave minerals than enter. A most likely process leading to high K/Rb ratios in all the granulite facies rocks is then one of "de-gassing" of deep crustal levels. The remaining stable minerals are of necessity depleted in Rb relative to K. The migrating disperse phase would be hydrous and enriched in K and Rb. It would contain other elements such as Th, rare earths, Y, Nb and U which are either incompatible in high grade minerals or easily leached from rocks as complexes in hydrous solutions. In view of the more basic nature of the granulite facies gneisses the migrating phase would also probably contain Si, Al, and Na. It will be shown that the egress of such a fluid from depth is necessary to accomplish the chemical changes seen in the succeeding retrogressive metamorphism.

Depletion in K and Rb at granulite facies then has given the observed high K/Rb ratios and also low K/Ba, and Rb/Sr and high Ba/Rb, Ba and Sr not being removed from the granulite facies. What is seen is a residue, partly purged of some elements and enriched in others. Taylor (1965) has noted that Ba and Sr are enriched in early fraction of igneous differentiation series. In this process of metamorphic fractionation they have been relatively enriched in the residuum. The presence of greater than normal crustal abundances of Ba and Sr in hypersthene-gneisses and basic granofelses is significant. Three possibilities exist to explain this feature. Firstly it could have been inherited from the original rocks. It might be due to massive depletion in other elements during the granulite facies so that the relative volumes of Ba and Sr have become larger with "basification" of the complex. Finally, if there has been a

de-gassing of the crustal level represented by the Tiree rocks, removing K and Rb, a similar process might have been operative at the base of the crust or in the upper mantle. In this the suite of elements incompatible with mineralogy might have been added to by Ba and Sr (Gast 1968). The diffusion of a disperse phase from depth, through the level exposed, could have brought Ba and Sr into an environment when they might leave their host fluid for surrounding felspars. In felspars they would be at an advantage to K and Rb in competing for sites, and displace the last two elements to the disperse phase, thereby adding a further boost to their expulsion.

At this point it is convenient to discuss the anomalous trend of K/Rb in metabasites and mafic granofels where Fig. 24b shows K/Rb decreasing from a maximum at about 0.8 per cent K to much lower values at low K contents. The only feasible explanations for this are either that some mineralogical feature at low concentrations of K and Rb favours the larger Rb^+ ion to K^+ in competition for some alien site, or at low levels of K in basic rocks, the calibration curve for x-ray analysis for K diverges from that used at Birmingham.

Chemical changes during retrogression.

Assuming common origins for all members of the various geochemically distinct groups of rocks, the profound differences with respect to certain chemical parameters between granulite and amphibolite facies members is persuasive to a hypothesis of a metamorphic fractionation process transforming the rocks during retrogression.

The main differences have been enumerated, and apparently the main changes have been:

a) an increase in K content in concordant granites, metabasites, granofelses, and ultramafic rocks. Little difference has been observed in K content of quartzo-felspathic gneisses and analysed metasediments.

b) An increase in Rb in all the rock types with consequent effects on K/Rb, Ba/Rb, and Rb/Sr ratios.

c) Much lower concentrations of Ba and Sr in migmatitic, nebulitic, diopside- and epidote-chlorite-gneisses, compared to hypersthene-gneisses.

Taylor's (1965) review of the behaviour of trace elements suggests various parameters whose variation may be used to follow the course of

chemical fractionation. The concentration of Ba in early K sites in igneous differentiation causes Ba to decrease with fractionation. The increase in K and Rb in successively later fractions causes the K/Ba ratio to increase and the Ba/Rb ratio to decrease rapidly with fractionation. In late stages of fractionation (Shaw's (1968) pegmatitic-hydrothermal stage) the effect of the larger ionic radius of Rb^{+} causes it to be enriched so that K/Rb ratios decrease from normal values. Sr^{2+} is captured in early K^{+} positions, and Sr decreases relative to K in fractionation with consequent increase in K/Sr. As Sr^{2+} favours 8-fold co-ordination sites to 6-fold Ca^{2+} sites Sr may be expected to increase relative to Ca in fractionation so that Ca/Sr ratios increase. The concentration of Sr in early K positions and the enrichment of Rb with fractionation results in an increase in Rb/Sr ratios.

The behaviour of the parameters K/Rb, K/Ba, Ba/Rb and Rb/Sr in the Lewisian rocks of Coll and Tiree is displayed by Figs. 24 and 25.

Examination of the variation of K/Ba with K (Fig. 24f) for concordant granitic veins shows a remarkable trend of increasing K relative to Ba. Although most plot as expected from the strong coherence of Ba with K (Fig. 19c) at reasonably constant K/Ba, several with large K contents have very large values of this ratio, up to 365. This is strong evidence for a metasomatic increase of K in the granites, correlated with the increase in microcline in deformed example and confirming the evidence of Q-Ab-Or and An-Ab-Or plots (Figs. 23a and 13c). The other two graphs, for supposed intrusive rocks and gneisses and metasediments, show similar but not such marked trends. For a range of K in granulite facies rocks, K/Ba varies little, which is probably a feature of their metamorphism, Ba tending to follow K and thus stabilising the ratio. There is a general trend of increasing K/Ba with increasing K in amphibolite facies rocks. It is best shown by ultramafic rocks where hornblende- and biotite-rich types have a high K/Ba and the four freshest have low ratios. The observed trends could be due to either K influx or Ba depletion during the amphibolite facies. In the metabasites there is no appreciable difference in Ba contents of granulite and amphibolite facies rocks, and the variation is probably due to the mobility of K. The trend in intermediate granofelses is scattered but similar. Unfortunately no comparable granulite facies rocks were found. For their high K contents, diopside-granofelses do not

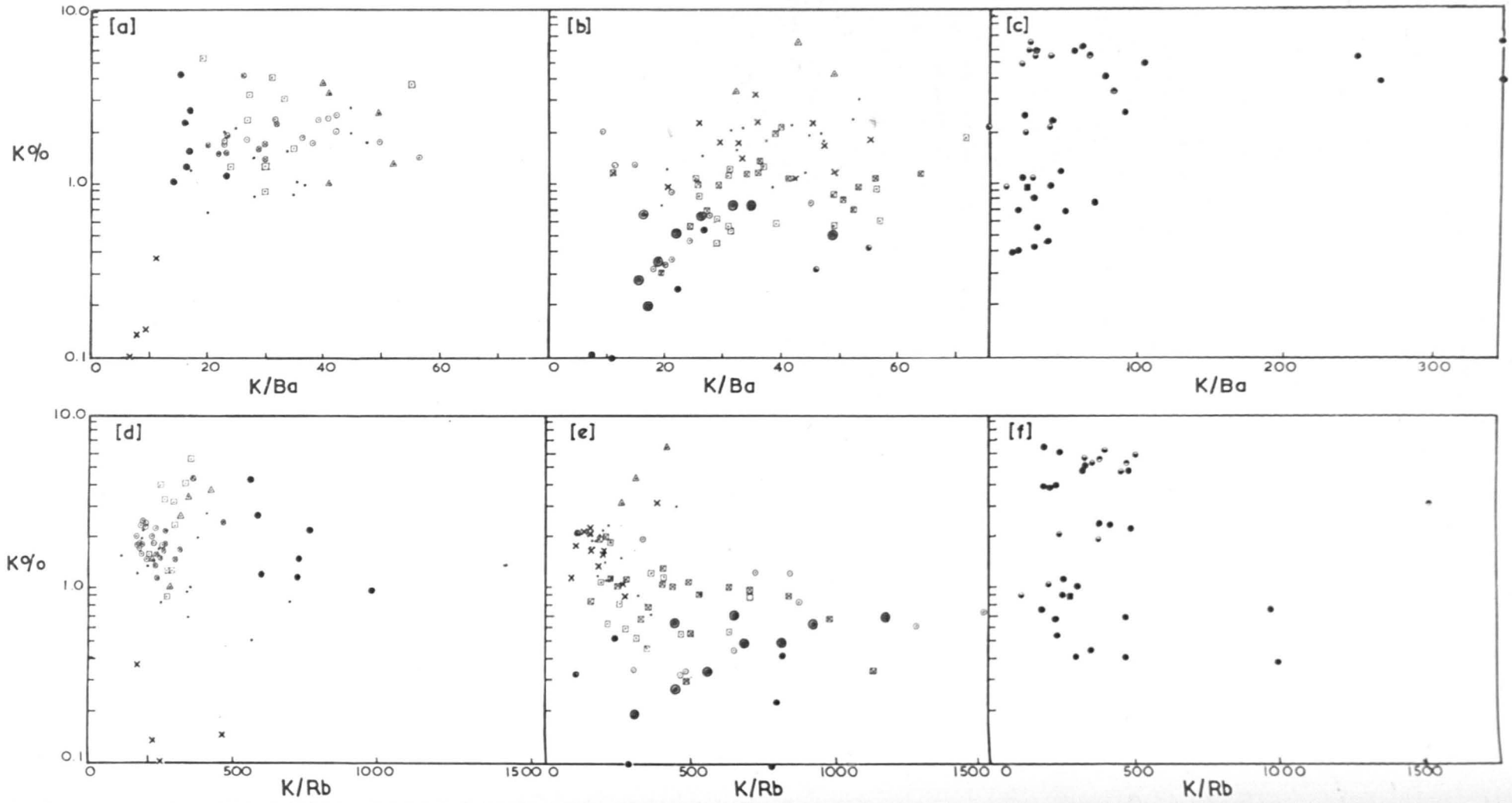


Fig. 24: graphs of logK against K/Ba and K/Rb.

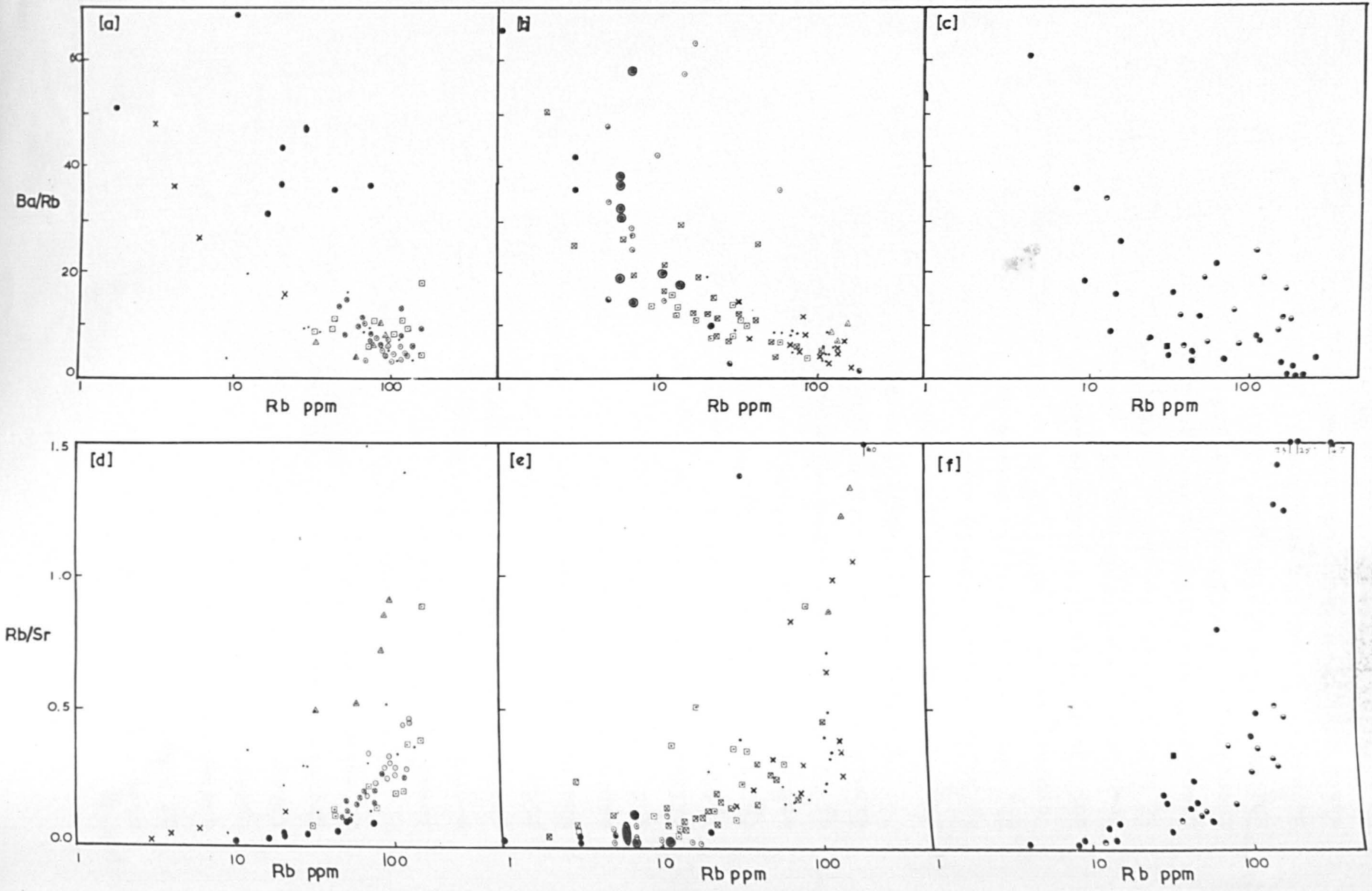


Fig.25: Graphs of Ba/Rb and Rb/Sr against log Rb.

show high K/Ba, perhaps as a result of Ba enrichment in their environment of strong mobilisation in diopside-gneisses. For a similar range of K, quartzo-felspathic gneisses show higher K/Ba in amphibolite facies representatives than in granulite facies, either as a result of Ba depletion (they have much lower abundances of Ba) or having been derived from K-poor granulite facies gneisses by K-metasomatism. The problem of Ba mobility is returned to later.

For a particular K content, amphibolite facies rocks have generally lower K/Rb ratios than granulite facies rocks (Fig. 24). There is also a trend as a whole towards decreasing K/Rb with increasing K content, probably best shown by the ultramafic rocks, where the outer zone of biotite-rich material has a much lower K/Rb and higher K than the fresh examples. Whatever the explanation for the K/Ba trends, either K-metasomatism or Ba depletion, in the amphibolite facies rocks Rb has increased relative to K and its influx into the system must be acknowledged. The slight but significant differences in K/Rb between migmatitic, nebulitic, diopside- and epidote-chlorite-gneiss are satisfactorily explained by their mineralogy (p. 105). Because of this in the event of Rb influx it would be preferentially accepted in biotite-rich migmatitic gneisses. The destruction of biotite in the sheared epidote-chlorite-gneisses may have led to Rb expulsion during greenschist facies metamorphism.

The variation of the ratios Ba/Rb and Rb/Sr with Rb are shown in Fig. 25. An increase in Rb content shows very well correlated decrease in Ba/Rb and increase in Rb/Sr. All granulite facies rocks, low in Rb, have very high Ba/Rb ratios and very low Rb/Sr ratios. The trends towards low Ba/Rb and high Rb/Sr are seen in the amphibolite facies rocks, and the increasingly deformed and potassic concordant granites. Apart from in the quartzo-felspathic gneisses, the Ba contents can be assumed to have remained constant, or in the case of the granitic veins to have increased with K, during retrogression. These trends then illustrate the increase in Rb relative to Ba and Sr, in all probability governed by Rb influx. Some significant effect may have been superimposed by decreasing Ba and Sr during the amphibolite facies. Now, the diopside-gneisses, which have been shown to have very low Sr and high Rb/Sr ratios compared to the other gneisses, plot on Rb/Sr against Rb (Fig. 25d) with a well defined trend parallel to that shown by the other gneisses. However it

is displaced towards high Rb/Sr for a particular Rb content. Thus although Rb has been introduced into these rocks, their Rb/Sr ratios have been amplified by loss of Sr, perhaps after Rb enrichment when they would have been on the main trend. The diopside-gneisses seem to have been formed at the same time as the nebulitic gneisses in a geochemically mobile environment. They are enriched in K, Mg and Ca possibly as a result of a metasomatic interchange at the nebulite/metasediment boundary. In response to the influx of Ca and Mg from the metasediments, Sr may have left, probably entering the metasediments or possibly entering fluids migrating out of the complex.

From the above discussion, it seems that the ratios K/Rb, Ba/Rb, K/Ba and Rb/Sr may be used, in this case, as quantitative measures of chemical fractionation in the retrogressive metamorphism. The degree and type of correlation coefficients of different trace elements with these parameters in correlation matrices for compositional groups may give some information as to the fractionation of elements. When considering Rb,Sr and Ba these can only be compared with validity against ratios from which they are absent, as their own variation will cause sympathetic variation in the ratios. In a similar fashion, any element showing strong correlation with an element appearing in a ratio, may show spurious correlation with that ratio. Although the method has pitfalls, probably deeper and more abundant than at first apparent, it would seem to be valid for studying such elements as P, Cu, Zn, Y, Zr, Nb, La, Ce, Pb and Th which rarely show any correlation with K, Rb, Sr, or Ba. Taking into consideration the drawbacks, some information may be obtained from Rb, Sr and Ba. It would be nice if another parameter were found to vary independantly of those listed, so that more meaningful information could be obtained from Rb, Sr and Ba variation.

The findings of a correlation study of this nature are given in Table 12. Only correlations significant above the 99 per cent level are shown. The study shows that in quartzo-felspathic gneisses, the fractionation process operative during retrogression has depleted these rocks in P_2O_5, Sr, and perhaps Ba, Th and Ca/Sr ratios increased with fractionation. Fig. 26 shows plots of Th, Sr and P_2O_5 against K/Rb for the gneisses. Concordant granitic veins show similar features but have apparently been depleted in La during retrogression perhaps due to the disappearance of

TABLE 12 : Student's t values for correlation coefficients

	K/Rb	K/Ba	Rb/Sr	Ba/Rb
Quartzo-felspathic gneisses, n = 50				
P_2O_5	5.45	-2.71	-3.95	5.02
Rb	-5.00	3.63	4.96	-4.83
Sr	8.54	-7.34	-7.51	10.47
Ba	2.77	-4.05	N.S.	3.15
Th	-4.20	N.S.	N.S.	-3.60
Metabasites, n = 41				
Pb	-2.96	N.S.	N.S.	-2.48
Granofelses, n = 37				
Sr	N.S.	-2.69	-2.39	3.48
Pb	-3.04	N.S.	N.S.	-3.00
Concordant granites, n = 26				
K_2O	N.S.	3.66	5.75	-2.65
P_2O_5	2.82	N.S.	N.S.	N.S.
Rb	N.S.	5.77	14.83	-2.9
Sr	2.42	-2.94	-3.32	2.71
La	5.03	N.S.	N.S.	5.26
Th	N.S.	2.66	N.S.	N.S.
Garnet-biotite-gneisses, n = 14				
La	N.S.	12.80	N.S.	N.S.

N.S. - not significant at 99 per cent level.

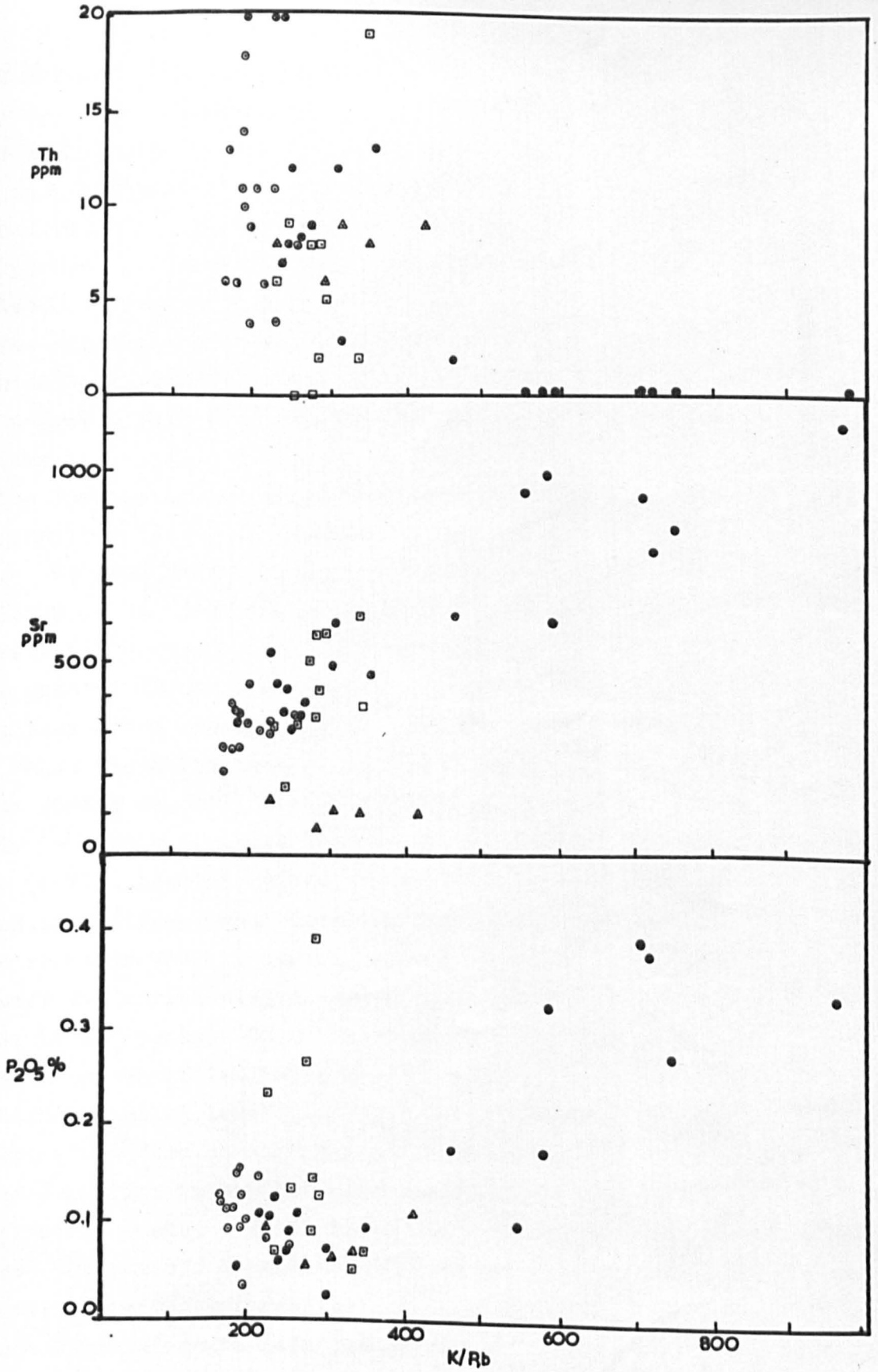


Fig.26: graphs of Th, Sr, and P_2O_5 against K/Rb for quartzo-felspathic gneisses.
Symbols as Fig.13a.

garnet. Metabasites and granofelses show a common increase in Pb with retrogression. The garnet-biotite-gneisses show increase in La with retrogression. Rb has been shown to have increased in the amphibolite facies rocks, as has K to a certain extent, and Ba has decreased in the gneisses. Therefore, although each of these elements either appears in the critical ratios or is positively correlated with other elements in the ratios, their correlations shown in Table 12 are partly significant.

The application of this method to much larger populations than those presented here would give more significant relationships, and might prove to be a useful tool in following the behaviour of trace elements in metamorphism.

The various mineralogical and chemical changes during the retrogression from granulite facies to amphibolite and greenschist facies have been noted. To produce the hydrous assemblages, H_2O must have been added to the system. Its addition is necessary to explain many facets of the deformational history. Any percolating hydrous fluid must have come from deeper crustal levels than those represented by the amphibolite facies unless they moved laterally from a high grade zone of de-gassing. This would necessitate the lateral migration of high grade metamorphic conditions during the long orogenic history. Neither situation can be positively resolved in the absence of contemporary higher grade regimes, either in the immediate vicinity or throughout the Lewisian. The de-gassing of the deep crust invoked for K and Rb depletion of the granulite facies rocks provides a source for the ionically enriched hydrous phase necessary for enrichening subsequently retrogressed rocks in K, Rb, Th, Pb, Ce, La and perhaps Cu. It would be instructive to consider the effects of such a medium of its being brought into contact with high grade rocks at a higher crustal level than that of their formation. Hydration of pyroxene would form hornblendes and biotite, influx of K would encourage biotite formation in the gneisses and the growth of microcline in very acid granitic rocks. Which trace elements leave or enter the migrating disperse phase would depend on their compatibility in the stable minerals and their concentration gradients across mineral boundaries. Rb and Th and some other elements left the disperse phase as shown, and Sr and Ba perhaps left the gneisses to enter the disperse phase and be carried to higher levels.

The last manifestations of the amphibolite facies are biotite-pegmatites. Their richness in rare earth elements, Zr, and Ba, high Ba/Sr and low Ca/Sr ratios are consistent with their being highly fractionated rocks. Their close similarity to deformed and remobilised concordant granites in many respects is persuasive to a close relationship between the two groups. An origin involving the effects of a residual portion of a diffusing disperse phase on the potassic granites may account for the peculiarities of the biotite-pegmatites. At a higher crustal level in the complex's history such fluids may have effected mobilisation of the concordant granites, either dissolving them or helping them to remelt. Such a process would explain both the highly fractionated nature, and similarity to the early granites, of the biotite-pegmatites.

It may be that what is now represented at the surface by the Lewisian rocks of Coll and Tiree is analogous to a chemical layering in the crust. Deepest levels are represented by the granulite facies rocks of Western Tiree, and decreasing depth represented by the retrogressed rocks. The evidence shows that the rocks have been metamorphosed at different grades since the granulite facies, and at each grade ~~has~~ have been subjected to the effects of migrating fluids. These fluids presumably originated at deeper levels, possible a proportion in the upper mantle. Throughout their migratory history they were probably in equilibrium with stable phases, and changed their composition appropriately. Elements incompatible in the metamorphic minerals at each level would tend to enter the disperse phase. Those elements able to compete at an advantage to incompatible elements for available sites would leave the migrating fluid to become stabilised in minerals under the physical conditions at particular levels. The process would, of necessity, be a continuous one, materials migrating from rocks at granulite facies finding their way into contemporary higher level, lower grade rocks. The inflowing fluid in amphibolite facies rocks must come from contemporary higher grade material, of deeper crustal levels. The coalesced fluids leading to the formation of late biotite-pegmatites may have been the end products of a long history of migration, and represent the latest portions of a metamorphic fractionation process acting through the intergranular medium of element transfer.

Such a hypothesis rests on the presence or absence of simple crustal

layering of metamorphic facies, and that during, say, the amphibolite facies phase the exposed rocks were underlain at some depth by contemporary granulite facies rocks providing the upwardly migrating fluids. Lambert and Heier (1968) come to similar conclusions, and envisaged the layer of high granitophile element (Si, K, Rb, U, Th, Pb) abundances in the low pressure granulite and amphibolite facies rocks of the Australian Shield to represent the crustal level at which elements, partially expelled from the lower crust during high grade metamorphism, have entered a stable environment. From data on medium to high pressure granulite facies rocks, they suggested that the lower crustal levels they represent have a more intermediate composition with low Th, U, Rb, Pb, Si and K, and higher abundances of Ca, Mg, Fe, Mn, and Ti, and are residual products of a process of metamorphic fractionation.

In contrast it is equally likely that the chemically contrasted events distinguished on Coll and Tiree were distinguished as much by the availability of migrating volatiles, or the ability of volatiles to migrate, as by depth controls, and that mobility of volatiles was influenced to a great extent by structure in the basement. Such an argument tends to be circular as it is well accepted that the presence of intergranular fluid phases controls deformation to some extent (e.g. Holland and Lambert, 1969; Orowan, 1965).

In the absence of concomitant granulite facies metamorphism below the amphibolite facies now represented by the bulk of the rocks on Coll, the migratory material of a hydrous and element enriched nature must have originated from the mantle or by lateral migration from metamorphic "highs" or possibly bell-shaped anomalies in the geothermal gradient. During the granulite facies metamorphism, in this case, little mantle-derived fluid can have entered the system, or else it passed through the rocks leaving them still anhydrous. In the amphibolite facies massive influx of volatiles must be envisaged. It is difficult to account for a situation in which "de-gassing" of the mantle could proceed sporadically, unless during the granulite facies, the anhydrous rocks provided a deep crustal barrier to migration, which "burst" at the onset of further deformation to allow volatiles to penetrate the formerly granulite facies rocks, and reduce these rocks, from the crust/mantle boundary to the accepted crustal level of amphibolite facies, to hydrous assemblages.

CHAPTER 5

CONCLUSIONS

To a large extent detailed conclusions, speculations, and discussions regarding the history of the Lewisian complex of Coll and Tiree are contained in the relevant chapters and their included sections. It remains here to attempt to relate the various aspects studied in order to give a brief history of this portion of the Lewisian and to note the implications. A vital inclusion is an effort to relate the Coll and Tiree sequences to the broad picture now emerging in the literature of the sequence of events characterising the Precambrian basement of Scotland. It was not the author's original wish to enter into regional correlation considerations in the field of Precambrian gneissose basement regimes in view of the controversies and well populated pitfalls characterising the problem at this time. However, in view of the almost "virgin" territory studied here, and its situation well to the south of hitherto described areas of Lewisian, it may be useful to make a few statements about its possible position in the Scottish Precambrian panorama. Table 13 gives a synopsis of the Lewisian history of the area.

On Coll and Tiree, as in the Outer Hebrides (Coward et al., 1969), metasediments and gneisses were present before granulite facies metamorphism. Coward and co-workers suggested that the weight of evidence pointed to an older basement covered by sediments. On Coll and Tiree distinctive metasediments, quartzites, calc-silicate rocks and marbles are in relatively narrow, continuous belts within the dominant gneisses. In Chapter 2 it was suggested that some of the belts on Coll might represent tightly appressed synforms, similar to those found in the Alps where Mesozoic sediments are in tight synforms separated by wide areas of Hercynian basement (Ramsay, 1967, Fig. 7-43). However, no structural evidence is present to separate these two components of the complex. A discussion on the possibilities of origin of the dominant gneisses is contained in Chapter 4, where it was concluded that no geochemical criteria can be used to definitely account for the present compositions found in the gneisses. It is worth restating that such aluminous rocks as garnet-

biotite-gneisses, which are commonly in thin bands in the dominant gneisses, are not necessarily metasediments but could have been derived by, for instance, element migration in mylonite zones during the pre-granulite facies period. The garnet-biotite-gneisses are chemically similar to others from South Harris and the Central block of the Mainland Lewisian. As with most of the other rocks, their chemistry, especially trace element abundances, is a reflection of their content of metamorphic minerals such as garnet.

At some stage after the deposition of the definite metasediments, the complex was intruded by basic igneous rocks, now represented by metabasites and mafic granofelses. The chemical similarity of all analysed metabasites from every metamorphic and structural environment suggests that they are all related, although many of their chemical characteristics are probably due to their hornblende-rich nature as Orville (1969) has suggested for the majority of metamorphic amphibolites. Their major element compositions are similar to tholeiitic basalts, although being very low in CaO. The mafic granofels differ in having normal CaO contents and high Ba and Sr contents. Certain variations seen in the metabasites, such as the negative correlations of Cr and Ni with TiO_2 and P_2O_5, and the positive correlation of MgO/Ni with SiO_2, point to their common origin as a differentiated series of basic igneous rocks. Others such as the positive correlations of Cr and Ni with MgO are controlled by probable metamorphic variation in hornblende content. There is a tendency for the larger metabasite bodies to be close to metasediment belts, but not all metasediment belts have associated large metabasite bodies. Coward et al. (1969) suggested that the close spatial relationships of metasedimentary rocks and basic bands could account for the preservation of metasediments, basic bands acting as barriers to element migration. In this case, such a relationship is not apparent. Metabasite intrusion may have been partly controlled by the sedimentary belts, or the larger metabasites could have been originally contemporaneous with sedimentation. Although the larger bodies' major element composition, similar to the bulk of narrower and discordant examples, need not rule out the latter possibility, their similarity in trace element chemistry makes it more likely that they are part of the same suite of intrusions.

As indicated at various points in this thesis, the lack of fabric in intermediate granofelses must be accounted for in discussing their origin. Their frequent association with metasediment belts led Survey workers (1930) to ascribe a sedimentary origin to them. That they can also occur in positions remote from sediment belts, and are more frequently associated with metabasites, has been shown in this study. In certain respects these granofelses are geochemically similar to metabasites in having negative correlation of Mg and Cr with TiO_2, but in other respects as on graphs of MgO/Ni against SiO_2 are not apparently members of an igneous differentiation series associated with metabasites. Their present compositions are probably more basic than originally, considering the large volumes of derived concordant granites within them. Some similarities with average analyses for andesites and dacites might suggest that they are either original andesitic-dacitic rocks present in the complex as an integral portion, or were intruded later. Any origin by igneous means might account for their present lack of fabric. A further, and equally likely possibility, is that they represent recrystallised, originally glassy, mylonitic rocks, developed during a pre-granulite facies phase. This might explain their frequent close associations with sediments and metabasites. The margins between quartzo-felspathic gneisses and metasediments, may at some stage have been a boundary between different bulk rheologies and a likely site for dislocation and mylonitisation. Mylonite bands in the complex may have encouraged metabasite intrusion to a certain extent. In Chapter 4, the similarity between intermediate granofelses and late flinty crush rock and pseudotachylite was noted. Like these particular glassy mylonitic rocks, the granofelses are more basic than surrounding gneisses. Obviously in the face of the recrystallised nature of these rocks and the lack of any strong evidence for origins, no choice can be made between possibilities as the basis of the methods applied.

Ultramafic rocks were emplaced in the complex at an early stage, later than the laying down of sediments as evidenced by the body in the Clabhach-Gorton belt on Coll. No relationships to metabasites have been seen and so their relative time of emplacement is unknown. That they were emplaced under metamorphic conditions is evidenced by their content of co-existing orthopyroxene and olivine. Their chemistry is similar to Lewisian ultramafic rocks on the Mainland, but is much different to Alpine-type

ultramafic rocks, containing high alkalies, Ba, Sr and Pb, perhaps due to their content of hornblende. Previously highly aluminous and calcic orthopyroxenes, now containing exsolved spinel and hornblende, indicates their presence in a high pressure-temperature regime before emplacement to higher levels. Their composition may represent mantle material, or may be due to influx of cations after emplacement. Their content of hornblende may be primary, although recrystallised, due to the effects of a contained hydrous phase in equilibrium with pyroxenes at depth and reacting with them on emplacement, or due to hydroxylation after emplacement. Their association with possible mylonitic rocks (garnet-biotite-gneisses) at Loch a' Phuill on Tiree may be indicative of their emplacement by tectonic means. Elsewhere they are present in metasediment belts, or isolated in country gneisses.

The presence of a foliation, folds and lineation in hypersthene-gneisses between Balephuil and Hynish on Tiree, relic sillimanite needles in deformed garnets in some garnet-biotite-gneisses with stable kyanite and the marked constancy of Fe/Mg ratios and Cr/Ni ratios in quartzo-felspathic gneisses is strong evidence of an orogenic history for the complex before granulite facies metamorphism. Discordant pyroxene-garnet-granitic rocks are common in hypersthene-gneisses and probably formed before the granulite-facies phase. They, together with the early foliation, are cut by metabasite dykes.

Concordant granitic veins in granofelses, metabasites and garnet-biotite-gneisses, frequently carrying garnet and sometimes pyroxenes, were formed before the granulite facies. Structural, petrographic and geochemical evidence are not compatible with their origin by intrusion, metasomatism or metamorphic differentiation, and only anatectic processes remain to account for their formation. Granitic rocks of pre-granulite facies age have described by Dearnley and Dunning (1968) from Adivachar Point on Benbecula, which are now deformed and have provided whole rock Rb/Sr ages of 2500 myr B.P. (Lambert et al., 1970). Sheraton (1970a) has described probable deformed granitic rocks, which he termed microcline-gneisses, from Scourian granulite facies terrain at Drumbeg.

Granulite facies rocks of all compositions are characterised by high K/Rb and Ba/Rb, and low K/Ba and Rb/Sr ratios. Concordant granites,

mafic granofelses and pyribolites have low Rb contents relative to normal values, and lower K content than amphibolite facies representatives. The evidence suggests that during the granulite facies metamorphism K, Rb, Y, Nb and Th were partially purged from the rocks. In considering the high K/Rb ratios, they are probably due to the incompatibility of Rb in cation sites in minerals stable at high grade. It is suggested that Rb, with other incompatible elements, was removed from the system, on breakdown of Rb enriched minerals such as micas, by a migrating intergranular fluid. This was contributed to by Rb expelled from stable minerals as a result of lattice collapse or the entry of elements such as Ba and Sr which are more favoured in substitution for various major elements because of their smaller ionic radii and stronger bond types. The abnormally high Ba and Sr contents of hypersthene-gneisses may have been caused by influx of the two elements in migrating fluids, originating at deeper levels, possibly within the upper mantle.

Such a phase of granulite facies metamorphism is found to have occurred in numerous localities in the Lewisian, noteably in the Central block of the Mainland where it has been termed the Scourian episode. Evans (1965), Holland and Lambert (1969), Sheraton (1970a,b) and Skinner (1970) have all established that during the Scourian, Rb was depleted relative to K, and K, Y and Th were depleted. It is reasonable to assume that the granulite facies event on Coll and Tiree can be equated with the Scourian. Moorbath et al. (1969) deduced from the fact that Lewisian rocks, on a graph of Pb^{207}/Pb^{204} against Pb^{206}/Pb^{204}, all fall beneath the primary growth curve for $U^{238}/Pb^{204} = 8.68$, on a straight line intercepting that curve at a point representing 2900 myr B.P., that U was depleted from all samples analysed 2900 myr ago. They equated this event with the Scourian granulite facies metamorphism, earlier Rb/Sr and K/Ar dates of around 2600 myr being attributed to later retrogressive events.

The presence of irregularly distributed hornblende, in equilibrium with pyroxenes, in many granulite facies rocks shows that dehydration of the complex did not proceed to completion. The presence of kyanite in aluminous rocks indicates that pressures were intermediate to high, as suggested by Green and Ringwood (1967). At a late stage during this episode an anhydrous reaction in basic rocks led to the formation of garnet, clinopyroxene and quartz, at the expense of orthopyroxene, plagioclase and ore. In oversaturated basic rocks, Green and Ringwood (1967)

suggest that such reactions may occur at constant pressure with waning temperature. The reaction does not seem to have been chemically controlled and indicates a change in physical conditions within the granulite facies.

The earliest tectonic activity which can be assigned to a particular event led to the formation of a penetrative foliation (S_1) banding, and intrafolial folds (F_1) in rocks of all compositions. The rheological properties of all lithologies seem to have been the same. Holland and Lambert (1969) suggest that such a destruction of previous structure, and common behaviour of all rock types is typical of deep crustal levels in the granulite facies, and is due to "hot creep". They have suggested that such deformation tends to lead to horizontal banding. Thus D_1 may have taken place during granulite facies.

D_2 has been shown to be characterised by contrasted viscosities, acid rocks behaving more competantly than basic rocks. It led to the production of folds of varying geometry, mainly buckle folds, along axial surfaces in acid rocks and similar fold styles in metabasites. Such a reversal of the usually observed viscosity contrast between acid and basic rocks may have been due to elevated temperatures during D_2, the relatively anhydrous nature of the complex at that time, or a combination of both.

Quartzose pegmatites can be precisely located in the tectonic sequence observed, and are post-D_2, pre-D_3. They bear a strong resemblance to pegmatitic bodies from Barra which cut pre-granulite facies dykes and are cut by dykes of the Scourie swarm. These have provided minimum whole rock Rb/Sr ages of 2500 myr B.P. (Dr. S. Moorbath, personal communication). Should the rocks from Coll and Tiree prove to be of the same age, both D_1 and D_2 would be Scourian tectonic events, D_2 occurring probably after the climax of metamorphism.

During D_3 deformation, which led to a simple extension in the plane of S_1 foliation, mafic bands in quartzo-felspathic gneisses behaved competantly, relative to the acid material, and were boudinaged. In some way the relative viscosity of the acid gneisses was lowered, and the most felsic types show the most profound effects of D_3 strain; i.e. the nebulitic gneisses. During D_3, which must have been a relatively long

phase of activity, a process of gneissification occurred, almost identical to the "late Scourian" episode of Harris described so lucidly by Myers (1970). A discussion of the visible effects is contained at the close of both Chapters 2 and 3. The main features to be noted here are that the culminating effect was the production of hugely deformed but almost completely recrystallised nebulitic gneiss, and a process of agmatisation was initiated in some metabasite bodies. In the nebulitic gneisses, agmatisation proceeded to its conclusion as a hornblendite ball carrying nebulite, by a process of deformation and ionic diffusion.

The cause of the D_3 deformation seems to have been the onset of activity of incoming hydrous fluids. Availability of H_2O caused the retrogression from granulite facies assemblages to those of the amphibolite facies in all but a few cases, mainly restricted to western Tiree and some thick metabasite and granofels bodies. It also probably lowered the viscosity of quartzo-felspathic gneisses resulting in their easy deformation, and consequent effects on the now relatively competant basic rocks. The D_3 phase is regarded as the major episode during which retrogression of the complex occurred at amphibolite facies.

Many interesting geochemical changes were associated with retrogression, which have been summarised at the close of Chapter 4. The main changes were: an increase in K content in concordant granitic veins, metabasites, granofelses, and ultramafic rocks; an increase in Rb in all rock types; much lower concentrations of Ba and Sr in quartzo-felspathic gneisses especially diopside-gneisses, compared with probably parental hypersthene-gneisses. Consequently all the rock types were given more normal element abundances, and K/Rb, Ba/Rb, and Rb/Sr ratios. The process effecting these changes is suggested to have been metamorphic fractionation involving the influx of a hydrous intergranular phase enriched in certain elements which had been removed from a deeper crustal level. Using the ratios K/Rb, Ba/Rb, Rb/Sr and K/Ba as indices of fractionation, it has been possible to show other changes during retrogression in the complex, which are adequately summarised by Table 12, the accompanying text in Chapter 4, and Fig. 26.

Amphibolite facies conditions seem to have continued during D_4 and D_5 whose properties are summarised in Chapter 2, Table 1. Both phases

had associated granitic fluid activity, manifesting itself as structurally controlled small granitic bodies. An interesting facet of the history of metamorphism in the amphibolite facies is the evidence for multiple phase growth of several mineral species. Each individual phase of formation of a particular mineral seems to have been differently initiated. Biotite shows this feature particularly well and has grown at the expense of orthopyroxene, garnet and green hornblende. Green hornblende has inverted from higher grade brown hornblende, replaced orthopyroxene via cummingtonite mantles, replaced clinopyroxene being epitaxically intergrown with it and resulting in quartz-sieved grains, finally a mixture of green hornblende, plagioclase and ore pseudomorphs earlier garnets in some amphibolites. The retrogressive phase is then seen to have been long and intricate.

No evidence of migmatisation, i.e. injection of numerous granite bodies and wholesale granitisation of gneisses was seen, after the process of gneissification in the retrogression phase was complete, as is seen in South Harris (Myers, 1970). The late discordant granitic rocks present are unlike those from South Harris and the Laxford area in being biotite- rather than muscovite-rich, nor do they have low K/Rb ratios. Their peculiar geochemistry has been described and it has been concluded that they are in some way related to earlier granitic material, but show some features indicating a highly fractionated portion included in them. Their similarity in structure and mineralogy to others widespread in the Lewisian is persuasive to their contemporaneity. Dr. R. St.J. Lambert (personal communication) has suggested that such rocks are distinctive and have a common Rb/Sr whole rock age of 2200 myr. It must be remembered that their present geochemistry may have been locally controlled and they may not be correlated with other like bodies. However, if they prove to be of the same age as Lambert suggests, this tends to confirm the observation that preceeding amphibolite facies gneissification during D_3 is of a similar age to that in Harris described by Myers (1970). Although Myers has preferred to refer to pre-dyke, post-granulite facies events as "late Scourian", many authorities (Evans, 1965; Sheraton 1970a; Evans and Lambert, in preparation) prefer the term Inverian for the period characterised by amphibolite facies metamorphism prior to intrusion of the Scourie dyke swarm ca. 2190 to 1900 myr ago (Evans and Tarney, 1964).

D_6 deformation, mainly in shear belts or ductile faults, gave rise to associated greenschist facies mineralogies. The main chemical effects on the gneisses was to increase their content of K and Ba, and to give some specimens anomalously low Ce/La ratios. This is suggestive, again, of the effects of migrating fluids of a highly fractionated nature. The absolute age of D_6 is unknown, but considering its low grade with respect to the Laxfordian of the Mainland and the Outer Isles, it is possibly post-Laxfordian; i.e. younger than 1750 myr (Lambert and Holland, 1970).

D_7 brittle structures have associated black flinty crush rock and epidotic crush rock. The chemistry of these materials indicates that even at the low grades of metamorphism associated with them (perhaps prehnite-pumpellyite metagreywacke facies) processes of ionic diffusion were operative. Flinty crush is more basic than parental migmatitic gneisses, and epidotic crush is highly calcic and aluminous. On Iona and the Rhinns of Islay the Geological Survey (1920) reported that epidote veins in the Lewisian are continuous with some in the overlying Torridonian. It is a distinct possibility therefore, that the late faulting on Coll with associated epidotic material is post-Lewisian.

In the Lewisian of Coll and Tiree an evolution can be traced from very early in this history of the complex, although the earlier the postulated processes, the less definite can any conclusions be. The main changes in the complex have always been geochemical, whether merely a redistribution of elements amongst minerals stable under prevailing physical conditions or wholesale migration of cations with consequent depletion or enriching of members of the complex in certain elements. In some cases the prevailing physical conditions have caused the geochemical activity, as in granulite facies metamorphism, in others the opposite has been true. Because of variation in temperature, pressure, and chemical composition (especially with respect to H_2O), structural features in the complex have changed with time, in terms of the effects of comparative viscosity in the complex, rates of strain and inhomogeneities in the rates of strain. These statements apply to all metamorphosed and deformed rocks.

Perhaps the most remarkable feature of the Lewisian of Coll and Tiree is the apparent change in metamorphic mineral facies from western Tiree

TABLE 13 : Lewisian history of Coll and Tiree

Tectonic sequence	Metamorphic conditions	Geochemical activity
I n d e c i p h e r a b l e e a r l y h i s t o r y		
At some stage complex consisted of sediments and parental material of gneisses.		
Early structures	Amphibolite facies possibly some mylonitisation.	Homogenisation of Fe/Mg and Cr/Ni ratios in hornblende and biotite in gneisses.
Emplacement of metabasites, mafic granofelses, ultramafic rocks.		
? SCOURIAN — INVERIAN — D_1	Upper amphibolite facies.	Partial melting of concordant granites.
D_2	Granulite facies.	Dehydration, depletion of K, Rb, Y, Nb, Th, possible influx of Ba, Sr in migrating phase from depth.
Emplacement of quartzose pegmatites.	Waning temperature, constant pressure.	
D_3	Amphibolite facies, garnet stable.	Influx of hydrous fluid, K, Rb, Th enrichment, Gneissification culminates in nebulitic gneiss and di-gneiss - redistribution of Ca, Mg, Rb, Sr. Agmatisation.
D_4	Garnet breakdown.	
D_5		
Emplacement of biotite-pegmatites.		Partial mobilisation of earlier granites and gneisses, encouraged by influx of highly fractionated hydrous fluid.
D_6	Greenschist facies.	Ba influx into gneisses, Ce/La decreases.
D_7	Prehnite-pumpellyite facies. Frictional fusion.	Mobility of many cations, giving flinty crush rock.
	Granulation.	Ca and Al concentrated in epidotic crush rock.

to the north-east of Coll, the highest grades in the former and the latest lower grades being best developed in the latter area. At an early stage in this thesis it was remarked that vertical or steeply dipping, N-S trending foliation occupies two thirds of the archipelago from west Tiree to the line of the Hogh Bay - Breachacha belt, a distance of some 30 km. Apparently no major zone of abrupt dislocation breaks the sequence as is the case with granulite and amphibolite facies terrain junctions elsewhere in the Lewisian. At the close of Chapter 4 it was reflected that the rocks of Coll and Tiree are metamorphically and chemically analogous to a simple metamorphic layering in the crust, and they have provided some interesting data on chemical variation in the processes of granulite facies metamorphism, and the retrogressive metamorphism of high grade rocks. They can provide data pertinent to the evolution of the crust in Shield areas, if a more rigorously controlled sampling is undertaken, perhaps on a grid basis, for the quartzo-felspathic gneisses especially. The presence of potassic rocks in all areas and at all grades would encourage radiometric dating to complement the relative history provided here.

REFERENCES

ALLSOP,H.L., NICOLAYSON,L.C. and HAHN-WEINHEIMER,P., 1969. Rb/K ratios and Sr isotopic compositions of minerals in eclogitic and peridotitic rocks. Earth Planet. Sci. Letters 5, 231-244.

BAKER,J.W., LEMON,G.G., GAYER,R.A. and MARSHMAN,R.R., 1968. The Dutch Gin Schists. Geol. Mag. 105, 493-494.

BAILEY,E.B. and ANDERSON,E.M., 1925. The geology of Staffa, Iona and Western Mull. Mem. Geol. Surv. Scot. 45.

BENNISON,G.M. and WRIGHT,A.E., 1969. The geological history of the British Isles. Edward Arnold, London.

BERTHELSEN,A., 1960. Structural studies in the Precambrian of Wesbern Greenland; part 2: geology of Tovq ussap nuna. Medd. om Gronland, bd. 123, nr. 1, 1-43.

BHATTACHARJEE,C.C., 1963. The late structural history and petrological history of the Lewisian rocks of the Meall Deise area north of Gairloch, Ross-shire. Trans. Geol. Soc. Glasgow 25, 31-60.

BOWES,D.R., 1967. The petrochemistry of some Lewisian granitic rocks. Min. Mag. 36, 342-363.

" , 1968a. An orogenic interpretation of the Lewisian of Scotland. Proc. 23rd Internat. Geol. Congr. 4,225-236.

" , 1968b. The absolute time scale and the subdivision of Precambrian roccks in Scotland. Geol. For. Stockh. Forh. 90, 175-188

" , 1969. The Lewisian of the North-West Highlands of Scotland. Amer. Assoc. Petrol. Geol. Memoir 12, 94-

" and GHALY,T.S., 1964. Age relations of Lewisian basic rocks south of Gairloch, Ross-shire, Scotland. Geol. Mag. 101, 150-160.

" and HOPGOOD,A.M., 1969. The Lewisian gneiss complex of Mingulay, Outer Hebrides, Scotland. Geol. Soc. Amer. Memoir 115, 317-360.

" and KHOURY,S.G., 1965. Successive periods of basic dyke emplacement in the Lewisian Complex, south of Scourie, Sutherland, Scott. J. Geol. 1, 295-299.

" and PARK,R.G., 1966. Metamorphic segregation banding in the Loch Kerry basite sheet from the Lewisian of Gairloch, Ross-shire, Scotland. J. Petrol. 7, 306-330.

" , WRIGHT,A.E. and PARK,R.G., 1964. Layered intrusive rocks in the Lewisian of the North-West Highlands of Scotland. Quart. J. Geol. Soc. Lond. 120, 153-192.

" " and " , 1966. Origin of ultrabasic and basic masses in the Lewisian. Geol. Mag. 103, 280-283.

BURNS,D.J., 1966. Chemical and mineralogical changes associated with the Laxfordian metamorphism of dolerite dykes in the Scourie - Loch Laxford area, Sutherland, Scotland. Geol. Mag. 103, 19-35

CARMAN,J.H. and TUTTLE,O.F., 1963. Experimental study bearing on the origin of myrmekite. Ann. Meeting Geol. Soc. Amer. 29A.

CHRISTIE,J.M., 1960. Mylonitic rocks of the Moine Thrust Zone in the Assynt district, North-West Scotland. Trans. Geol. Soc. Edinb. 18, 79-93.

CONDIE,K.C., 1967. Geochemistry of early Precambrian graywackes from Wyoming. Geochim. Cosmochim. Acta 31, 2135-2149.

COOMBS,D.S., 1960. Lower grade mineral facies in New Zealand. Proc. 21st Int. Geol. Congr. 13, 339-351.

COWARD,M.P., FRANCIS,P.W., GRAHAM,R.H., MYERS,J.S. and WATSON,J., 1969. Remnants of an early metasedimentary series in the Lewisian Complex of the Outer Hebrides. Proc. Geol. Assoc. 80, 387-408.

CRUFT,E.F., 1966. Minor elements in igneous and metamorphic apatites. Geochim. Cosmochim. Acta 30, 375-398.

CURRIE,K.L., 1968. On the stability of albite in supercritical water in the range 400° to 600°C and 750 to 3500 bar. Amer. J. Sci. 266, 321-341.

DAVIDSON,C.F., 1943. The Archaean rocks of the Rodil district, South Harris, Outer Hebrides. Trans. Roy. Soc. Edin. 61, 71-112.

DEARNLEY,R., 1962. An outline of the Lewisian complex of the Outer Hebrides in relation to that of the Scottish Mainland. Quart. J. Geol. Soc. Lond. 118, 143-176.

" , 1963. The Lewisian complex of South Harris; with some observations on the metamorphosed basic intrusions of the Outer Hebrides, Scotland. Quart. J. Geol. Soc. Lond. 119, 243-312.

" and DUNNING,F.W., 1968. Metamorphosed and deformed pegmatites and basic dykes in the Lewisian complex of the Outer Hebrides, and their geological significance. Quart. J. Geol. Soc. Lond. 123 353-378.

DE WAARD,D., 1965. The occurrende of garnet in the granulite facies terrane of the Adirondack Highlands. J. Petrol. 6, 165-191.

" , 1966. The biotite-cordierite-almandine subfacies of the hornblende-granulite facies. Canadian Mineralogist 8, 481.

" , 1967. The occurrence of garnet in the granulite facies terrane of the Adirondack Highlands, and elsewhere; an amplification and a reply. J. Petrol. 8, 210-232.

DEWEY,J.F., 1965. Nature and origin of kink-bands. Tectonophysics 1, 459-494.

DIETRICH,R.V. and MEHNERT,K.R., 1960. Proposal for the nomenclature of migmatites and associated rocks. Proc. 21st Internat. Geol. Congr. 26, 56-67.

EVANS,C.R., 1965. Geochronology of the Lewisian basement near Lochinver, Sutherland. Nature 207, 54-56.

" and PARK,R.G., 1965. Potassium-argon age determinations from the Lewisian of Gairloch, Ross-shire, Scotland. Nature 205, 350-352.

" and TARNEY,J., 1964. Isotopic ages of Assynt dykes. Nature 204, 638-641.

FAHRIG,W.F. and EADE,K.E., 1968. The chemical evolution of the Canadian Shield. Can. J. Earth Sci. 5, 1247-1252.

FLEUTY,M.J., 1964. The description of folds. Proc. Geol. Assoc. 75, 461-492.

FLINN,D., 1967. The metamorphic rocks of the southern part of the mainland of Shetland. L'pool Manch. Geol. J. 5, 251-290.

GAST,P.W., 1968. Trace element fractionation and the origin of tholeiitic and alkaline magma types. Geochim. Cosmochim. Acta 32, 1057-1086.

GHALY,T.S., 1966. The Lewisian geology of the area between Loch Shieldaig and Loch Braigh Horrisdale, Gairloch, Ross-shire. Scott. J. Geol. 2, 282-305.

GILETTI,B.J., MOORBATH,S. and LAMBERT,R.ST.J., 1961. A geochronological study of the metamorphic complexes of the Scottish Highlands. Quart. J. Geol. Soc. Lond. 117, 233-264.

GOLDSMITH,R., 1959. Granofels, a new metamorphic rock name. J. GEOL. 67, 109-110.

GOLES,G.G., 1967. Trace elements in ultramafic rocks. In: Wyllie,P.J. Ed., Ultramafic and related rocks. Wiley, New York.

GOODWIN,A.M., 1968. Archaean protocontinental growth and early crustal history of the Canadian Shield. Proc. 23rd Internat. Geol. Cong. 1, 69-89.

GREEN,D.H., 1964. The petrogenesis of the high temperature peridotite intrusion in the Lizard area, Cornwall. J. Petrol. 5, 134-188.

" and RINGWOOD,A.E., 1967. An experimental investigation of the gabbro to eclogite transformation and its petrological applications. Geochim. Cosmochim. Acta 31, 767-833.

GRESENS,R.L., 1966. The effect of structurally produced pressure gradients on diffusion in rocks. J. Geol. 74, 307-321.

GRIFFIN,W.L., and HEIER,K.S., 1969. Parageneses of garnet in granulite facies rocks, Lofoten - Westeraalen, Norway. Contr. Mineral. and Petrol. 23, 89-116.

HART,S.R., 1964. Ultramafic rocks of St. Paul's Island. Carnegie Inst. Wash. Yearbook 63, 330-331.

" , 1965. Potassium, rubidium and strontium in the ultramafic rocks of St. Paul's Islands. Geol. Soc. Amer. Spec. Paper 82, 86-87 (Abstr.).

" ,ALDRICH,L.T., TILTON,G.R., DAVIS,K.L., KROGH,T.E. and YAMAGUCHI, M., 1965. Potassium - rubidium studies of ultramafic rocks in Japan. Carnegie Inst. Wash. 64, 293-296.

HASKIN,L.A., FREY,F.A., SCHMIDT,R.A. and SMITH,R.H., 1966. Meteoritic, solar and terrestrial rare earth distributions. Phys. Chem. of the Earth 7, 167-323.

HEIER,K.S., 1960. Petrology and geochemistry of high grade metamorphic and igneous rocks on Langøy, Northern Norway. Norges Geol. Unders. 207, 1-246.

" , 1965. Metamorphism and the chemical differentiation of the crust. Geol. For. Stockh. Forh. 87, 249-256.

" and ADAMS,J.A.S., 1964. The geochemistry of the alkali metals. Phys. Chem. of the Earth 5, 253-381.

HIMMELBERG,G.R. and PHINNEY,WM.C., 1967. Granulite facies metamorphism. Granite Falls - Montevideo area, Minnesota. J. Petrol. 8, 325-348.

HITCHON,B., 1960. The geochemistry, mineralogy andorigin of pegmatites from three Scottish Precambrian metamorphic complexes. Proc. 21st Internat. Geol. Cong. 17, 36-52.

HOLLAND,J.G. and LAMBERT,R.ST.J., 1969. Structural regimes and metamorphic facies. Tectonophysics 7, 197-217.

HOLLAND,T.H., 1900. The petrology of Job Charnock's tombstone. J. Asiatic Soc. Bengal 62, 162-164.

HOLMES,A., 1920. Nomenclature of Petrology. Murby, London.

HOWIE,R.A., 1965. The pyroxenes of metamorphic rocks. In: Pitcher, W.A. and Flinn,G.W., Eds., Controls of Metamorphism; Oliver and Boyd, London and Edinburgh.

HUBBARD,F.H., 1965. Antiperthite and mantled felspar textures in Charnockite (enderbite) from S.W. Nigeria. Am. Miner. 50, 2040.

JEHU,T.R. and CRAIG,R.M., 1923. Geology of the Outer Hebrides. Part I Trans. Roy. Soc. Edin. 53, 419-441.

" and " , 1925. Geology of the Outer Hebrides. Part II. Trans. Roy. Soc. Edin. 53,615-641.

" and " , 1926. Geology of the Outer Hebrides. Part III. Trans. Roy. Soc. Edin. 54, 467-489.

JEHU,T.R. and CRAIG,R.M., 1934. Geology of the Outer Hebrides. Part IV. Trans. Roy. Soc. Edin. 57, 839-874.

KAMP, VAN DE, P.C., 1967. Geochemistry of metasediments in the Haliburton - Madoc area, South-East Ontario. Can. Jour. Earth Sci.5,1337-1372.

KANISAWA,S., 1969. Garnet-amphibolites at Yokohama, in the Abukuma metamorphic belt, Japan. Contr. Mineral. Petrol. 20, 164-176.

KHOURY,S.G., 1968a. The structural geometry and geological history of the Lewisian rocks between Kylesku and Geigeil, Sutherland, Scotland. Krystalinikum, 6.

" , 1968b. Structural analysis of complex fold belts in the Lewisian north of Kylesku, Sutherland, Scotland. Scott. J. Geol. 4, 109-120.

KING,B.C., 1965. The nature and origin of migmatites: metasomatism or anatexis. In: Pitcher, W.S. and Flinn G.W., Eds., Controls of Metamorphism, Oliver and Boyd, London and Edinburgh.

KOLBE,P. and TAYLOR,S.R., 1966. Major and trace element relationships in granodiorites and granites from Australia and South Africa. Contr. Mineral. Pe trol. 12, 202-222.

KRETZ,R., 1959. Chemical study of garnet, biotite and Hornblende from gneisses of S.W. Quebec, with emphasis on the distribution of elements in coexisting minerals. J. Geol. 67, 371-402.

LAMBERT,I.B. and HEIER,K.S., 1969. Geochemical investigations of deep seated rocks in the Australian Shield. Lithos 1 , 30-53.

LAMBERT,R.ST.J. and HOLLAND,J.G., in press. A chronological study of the Lewisian of the Laxford area, north-west Scotland. Abstract in Geol. Soc. Lond. Circular 158.

LAMBERT,R.ST.J., EVANS,C.R. and DEARNLEY,R., 1970. Isotopic ages of dykes and pegmatitic gneiss from the southern islands of the Outer Hebrides. Scott. J. Geol. 6, 208-213.

LEAKE,B.E., 1964. The chemical distinction between ortho- and para-amphibolites. J. Petrol. 5, 238-254.

" , 1969. A catalog of analysed calciferous and subcalciferous amphiboles, together with their nomenclature and associated minerals. Geol. Soc. Amer. Spec. Paper 98.

MACGREGOR,I.D., 1967. Mineralogyof model mantle compositions. In: Wyllie, P.J.,Ed., Ultramafic and related rocks, Wiley, New York.

MACLEAN,D., 1965. Science of metamorphism in metals. In: Pitcher,W.S. and Flinn, G.W., Eds., Controls of Metamorphism, Oliver and Boyd, London and Edinburgh.

MEHNERT,K.R., 1968. Migmatites and the origin of granitic rocks. Elsevere, Amsterdam.

MISCH,P., 1968. Plagioclase compositions and non-anatectic origin of migmatitic gneisses in northern Cascade Mountains of Washington State. Contr. Mineral. Petrol. 17, 1-70.

MOORBATH,S., STEWART,A.D., LAWSON,D.E. and WILLIAMS,G.E., 1967. Geochronological studies on the Torridonian sediments of North-West Scotland. Scott. J. Geol. 3, 389-412.

" , WELKE,H. and GALE,N.H., 1969. The significance of lead isotope studies in ancient high grade metamorphic basement complexes, as exemplified by the Lewisian rocks of Northwest Scotland. Earth Planet Sci. Letters 6, 245-256.

MOXHAM,R.L., 1965. Distribution of minor elements in coexisting hornblendes and biotites. Can. Miner. 8, 204-240.

MUELLER,R.F., 1961. Analysis of relations among Mg, Fe and Mn in certain metamorphic minerals. Geochim. Cosmochim. Acta 25,267-296.

" , 1967. Mobility of the elements in metamorphism. J. Geol. 75, 565-582.

MURTHY,V.R. and STUEBER,A.M., 1967. Potassium/rubidium ratios in mantle derived rocks. In: Wyllie,P.J., Ed. Ultramafic and related rocks, Wiley, New York.

MYERS,J.S., 1970. Gneiss types and their significance in the repeatedly deformed and metamorphosed Lewisian Complex of Western Harris, Outer Hebrides. Scott. J. Geol. 6, 186-199.

NOCKOLDS,S.R., 1966. The behaviour of some elements during fractional crystallisation of magma. Geochim. Cosmochim. Acta 30, 267-278.

ODE,H., 1960. Faulting as a velocity discontinuity in plastic deformation. In: Griggs,D. and Handin,J., Eds., Rock deformation, Geol. Soc. Amer. Mem. 79.

O'HARA,M.J., 1961. Zoned ultrabasic and basic igneous masses in the early Lewisian metamorphic complex at Scourie, Sutherland. J. Petrol. 2, 248-276.

" , 1962. Some intrusions in the Lewisian complex near Badcaul, Sutherland. Trans. Geol. Soc. Edinb. 19, 201-207.

" , 1965. Origin of ultrabasic and basic masses in the Lewisian. Geol. Mag. 102, 296-314.

" , 1967a. Mineral facies in ultrabasic rocks. In: Wyllie,P.J., Ed., Ultramafic and related rocks, Wiley, New York.

" , 1967b. Mineral parageneses in ultrabasic rocks. In: Wyllie,P.J., Ed., Ultramafic and related rocks, Wiley, New York.

OROWAN,E., 1965. Convection in a non-Newtonian mantle, continental drift and mountain building. Phil. Trans. Roy. Soc.Lond. Ser.A,258,284-313.

ORVILLE,P.M., 1969. A model for metasomatic origin of thin layered amphibolites. Amer. J. Sci. 266 542.-560

OSBOURNE,E.F., 1969. The complimentariness of orogenic andesite and alpine peridotite. Geochim. Cosmochim. Acta 33, 307-324.

OXBURGH,E.R., 1964. Petrological evidence for the presence of amphibole in the upper mantle and its petrogenetic and geophysical implications. Geol. Mag. 101, 1-19.

PARK,R.G., 1961. The pseudotachylite of the Gairloch district. Amer. J. Sci. 259 542-550

" , 1964. The structural history of the Lewisian rocks of Gairloch, Wester Ross, Scotland. Quart. J. Geol. Soc. Lond. 120, 397-433.

" , 1965. Early metamorphic complex of the Lewisian north-east of Gairloch, Ross-shire, Scotland. Nature 207, 66-68.

" , 1966. Nature and origin of Lewisian basic rocks of Gairloch, Ross-shire. Sctt. J. Geol. 2, 179-199.

PARRAS,K., 1958. On the charnockites in the light of a highly metamorphosed rock complex in Southwest Finland. Bull. Comm. Geol. Fin. 29, 1-37.

PASHLEY,W.D., 1956. The study of epitaxy in thin surface films. Adv. Phys. 5, 173.

PEACH,B.N., HORNE,J., GUNN,W., CLOUGH,C.T. and HINXMAN,L.W., 1907. The geological history and structure of the North-West Highlands of Scotland. Mem. Geol. Surv. Great Britain.

PHILLPOTTS,J.A. and SCHNETZLER,C.C., 1970. Phenocryst - matrix partition coefficients for K, Rb, Ba, and Sr, with applications to anorthosite and basalt genesis. Geochim. Cosmochim. Acta 34, 307-322.

PICHAMUTHU,C.S., 1959. The charnockite problem. Proc. Mysore Geol. Assoc..

POWER,G.M. and PARK,R.G., 1969. A chemical study of five amphibolite bodies from the Lewisian of Gairloch, Ross-shire. Scott. J. Geol. 5,26-41.
RABBITT,J.C., 1948. Anthophyllite in gneisses and amphibolites. Amer. Miner. 33, 263.
RAMBERG,H., 1952. The origin of metamorphic and metasomatic rocks. University of Chicago press.
" , 1956. Pegmatites in West Greenland. Geol. Soc, Amer. Bull. 67, 185-213.
RAMSAY,J.G., 1967. Folding and Fracturing of rocks. McGraw-Hill, New York.
" and GRAHAM,R.H., 1970. Strain variation in shear belts. Can. J. Earth Sci. 7, 786-813.
RAST,N., 1965. Nucleation and growth of metamorphic minerals. In: Pitcher, W.S. an Flinn, G.W., Eds., Controls of Metamorphism, Oliver and Boyd, London.
RICHIE,J.E. and THOMAS,H.H., 1930. The geology of Ardnamurchan, N.W. Mull and Coll. Mem. Geol. Surv. Scot. 51.
SAXENA,S.K., 1969. Analysis of equilibria involving garnet in rocks of granulite facies: a discussion. Amer. J. Sci. 542-
SEN,S.K., 1959. Potassium content of natural plagioclase and the origin of antiperthites. J. Geol. 67, 479-495.
SHAW,D.M., 1968. A review of K - Rb fractionation trends by covariance analysis. Geochim. Cosmochim. Acta 32, 573-601.
SHELLEY,D., 1967. Myrmekite and myrmekite-like intergrowths. Miner. Mag. 36,491-503.
SHERATON,J.W., 1970a. Unpublished Ph.D. thesis, University of Birmingham.
" , 1970b. The origin of the Lewisian gneisses of northwest Scotland, with particular reference to the Drumbeg area, Sutherland. Earth Planet. Sci. Letters 8, 301-310.
SIGHINOLFI,G.P., 1969. K/Rb ratios in high grade metamorphism: a confirmation of the hypothesis of a continual crustal evolution. Contr. Mineral. Petrol. 21, 346-356.
SINCLAIR,I.G.L., 1964. Unpublished Ph.D. thesis University of St. Andrews.
" , 1966. Amygdaloidal texture in the Lewisian rocks of the island of Tiree. Geol. Mag. 103, 336-339.
SKINNER,A.C., 1970. Unpublished Ph.D. thesis University of Birmingham.
SPRY,A., 1969. Metamorphic textures. Pergamon press, London.
SUBRAMANIAM,A.P., 1959. Charnockites of the type area near Madras - a reinterpretation. Amer. J. Sci. 257, 321-342.
SUTTON,J. and WATSON,J., 1951a. The pre-Torridonian history of the Loch Torridon and Scourie areas, and its bearing on the chronological history of the Lewisian. Quart. J. Geol. Soc. Lond. 106, 241-307.
" and " , 1951b. Varying trends in the metamorphism of dolerite dykes. Geol. Mag. 88, 25-35.
" and " , 1959. Metamorphism in deep seated zones of transcurrent movement at Kungwe Bay, Tanganyika Territory. J. Geol. 67, 1-13.
" and " , 1962. Further observations on the margin of the Laxfordian complex of the Lewisian near Loch Laxford, Sutherland. Trans. Roy. Soc. Edin. 65, 89-106.
TARNEY,J., 1963. Assynt dykes and their metamorphism. Nature 199, 672-674
" , 1964. Unpublished Ph.D. thesis University of Durham.
TAYLOR,S.R., 1964. Abundance of chemical elements in the continental crust: a new table. Geochim. Cosmochim. Acta 28, 1273-1285.
" , 1965. The application of trace element data to problems in petrology. Phys. Chem. of the Earth 6, 133-213.

TAYLOR,S.R., CAPP,A.C. and GRAHAM,A.L., 1969. Trace element abundances in andesites. II Saipan, Bougainville and Fiji. Contr. Mineral. Petrol. 13, 1-26.

" and WHITE,A.J.R., 1965. Geochemistry of andesites and growth of continents. Nature 208, 271-273.

" and " , 1966. Trace element abundances in andesites. Bull. Volcan. 29, 177-194.

THORPE,R.S., 1970. Unpublished Ph.D. thesis University of Birmingham.

TURNER,F.J., 1968. Metamorphic petrology. McGraw-Hill, New York.

" and VERHOOGEN,J., 1960. Igneous and metamorphic petrology. McGraw-Hill, New York.

" and WEISS,L.E., 1964. Structural analysis of metamorphic tectonites. McGraw-Hill, New York.

WAGER,L.R. and MITCHELL,R.L., 1951. The distribution of trace elements during strong fractionation of basic magma - a further study of the Skaergaard intrusion, East Greenland. Geochim.Cosmochim. Acta 1,129-208.

WASHINGTON,H.S., 1922. Deccan traps and other plateau basalts. Geol. Soc. Amer. Bull. 33,765-804.

WATSON,J., 1969. The Precambrian gneiss complex of Ness, Lewis, in relation to the effects of Laxfordian regeneration. Scott.J. Geol.5, 269-285.

WENTWORTH,C.K. and WINCHELL,H., 1947. Koolau basalt series, Oahu, Hawaii. Geol. Soc. Amer. Bull. 58, 49-78.

WHEATLEY,T.J., 1968. Unpublished Ph.D. thesis University of Birmingham.

WHITE,A.J.R., 1966. Genesis of migmatites from the Palmer region of South Australia. Chem. Geol. 1, 165-200.

WHITNEY,P.R., 1969. Variations in K/Rb ratios in migmatitic paragneisses of the Northwest Adirondacks. Geochim. Cosmochim. Acta 33, 1023-1211.

WILKINSON,S.B., 1907. The geology of Islay. Mem. Geol. Surv. Scot. 19 & 27.

WINKLER,H.G.F., 1967. Petrogenesis of metamorphic rocks. Springer Verlag, Berlin.

YODER,H.S. and TILLEY,C.E., 1962. Origin of basalt magmas: an experimental study of natural and synthetic rock systems. J.Petrol. 3,342-532.

APPENDIX 1 : GEOCHEMICAL METHODS

Sample collection

Approximately 1 kg samples were collected representative of the lithological variation seen. No attempt was made to collect a suite of samples volumetrically representative of the proportions of different lithologies, the proportion being impossible to determine in the field. The heterogeneity of the lithology precluded meaningful grid sampling.

Sample preparation

Each rock sample was initially split into 5 cm pieces in a hydraulic rock-splitter. These were then crushed on a tungsten carbide plate in a flypress with a tungsten carbide pestle until all fragments were finer than 5 mm. This material was ground in batches, in the tungsten carbide barrel of a "Tema" swing mill for 30 seconds per batch. The resulting powder was repeatedly quartered to a representative sample of about 100 gm which was ground for a further 3 minutes in the swing mill. To 7 gm of the resulting fine powder was added 14 to 18 drops of a solution of "Mowiol" (polyvinyl alcohol) in a 1:5 mixture of ethanol and water, and a paste was made. The paste was placed between highly polished tungsten carbide plattens in a vacuum die and pressed in a hydraulic ram at a jaw load of 15 tons. The pressed-paste was retrieved as a pellet 3 cm in diameter and about 1 cm thick, and allowed to dry.

In some cases, the bulk of coarsely ground rock was sieved, fractions between B.S.A. 100 and 150, and 150 and 200 mesh being washed and dried prior to mineral separation. This was carried out using a Frantz isodynamic magnetic separator, the rock being devided into quartz and felspar and various fractions of the ferromagnesian minerals present. Quartz-felspar mixtures were purified using tetrabromethane, thereby removing contaminants of ferromagnesion minerals, and heavy non-magnetic minerals. Each ferromagnesian mineral separated was purified using Clerici solution. The purified materials were ground in the swing mill. Quartz-felspar mixtures were pelleted in the normal way. Ferromagnesian minerals were mixed in the accurately weighed proportion 2 gm mineral to 4 gm "Spectroflux",

the mixed powder being fused at 800°C in "Specpure" carbon crucibles. The resulting glasses were ground and pelletted in the fashion described to ensure that inhomogeneities in the glasses were eliminated. Hornblendes were not fused, but pressed as powder pellets as precise trace element analyses were needed, at suspected low concentrations.

X-ray fluorescence analysis

26 elements were determined on the departmental PW 1212 XRF spectrometer. Details of the principles of X.R.F. spectrometry may be found in Jenkins and de Vries (1967), use of PW 1212 in Phillips manual Scientific Equipment 93 737 59. 1y. 10, and the applications of X.R.F. techniques to rock analysis are well covered by Leake et al. (1969).

As the instrument used automatically analysed x-rays on 15 channels with three tubes, with Cr, W and Mo targets, the suites of elements was analysed by 5 separate machine programs. Machine conditions for each of the programs are summarised in Table 14.

Calibration for rock analyses was carried out essentially as described by Leake et al. (1969), approximately 180 analysed rocks of acid, intermediate, basic and ultrabasic igneous nature, and comparable metamorphic rocks being used to plot calibration curves for each major element. The initial calibration curves (concentration of element versus counts relative to 1000 counts on a ratio standard) were constructed at the University of Birmingham by Drs. J. Tarney and G.L. Hendry, and Messrs. A.C. Skinner, J.W. Sheraton and T.J. Wheatley. Further contributions have been made by other members of the Geology Department. All the calibration curves with the exception of those for Al and K where quadratic relationships were found, were linear. Trace element calibrations were effected by a "spiking" or standard addition technique.

It was found that different calibration coefficients were obtained from different compositional groups of rocks, presumably as a result of variations in mass absorption coefficients. The different calibration coefficients were built into the machine program as arbitrary block codes, depending on SiO_2 and total Fe content of the rocks, described by Skinner (1970).

In a few cases count rates for particular element peaks had to be corrected for interference by peaks of other elements, for instance

TABLE 14: Operating conditions for PW 1212 X.R.F.S..

Chromium tube programs

	Program "Cr"						Program "L"	
Element	Ti	Ca	K	Si	Al	P	Mg	Na
Peak	K_α	K_α	K_α	K_α	K_α	K_α	K_α	K_α
Bkgd.°2θ	+3.20	-1.80	+3.75	+3.00	-4.80	+2.70	-2.54	+1.05
kV	40	20	20	20	60	60	50	50
mA	16	4	16	32	24	32	40	40
Coll.( )	160	480	480	480	480	480	480	480
Crystal	PE	PE	PE	PE	PE	Ge	KAP	KAP
Window( )	3.75	3.75	3.75	3.75	3.75	3.75	2.0	2.0
Set cnts.	1×10^5	1×10^5	1×10^5	3×10^5	3×10^5	-	-	-
Time,peak	-	-	-	-	-	40	40	100
Time,bkgd.	-	-	-	-	-	40	20	40

Gas flow proportional counter and vacuum path throughout.

Tungsten tube program

	Program "W"					
Element	Ni	Fe	Mn	Cr	Ce	Ba
Peak	K_α	K_α	K_α	K_α	L_{β_1}	$L_{\beta_{1,4}}$
Bkgd.°2θ	WLL	+4.83	-1.70	-2.22	-1.75	+2.69
kV	60	20	60	60	60	60
mA	32	16	32	32	32	32
Coll.( )	160	160	480	480	480	480
Detector	GFPC	GFPC	Both	GFPC	GFPC	GFPC
Set cnts.	1×10^5	3×10^4	1×10^4	1×10^4	1×10^4	3×10^4

LiF (220) crystal, 3.75 window, vacuum path throughout.

	Tungsten tube program						Molybdenum tube program	
	Program "T"						Program "Mob"	
Element	Zr	Y	Sr	Rb	Th	Pb	Zn	Cu
Peak	K_α	K_α	K_α	K_α	L_{α_1}	$L_{\beta_{1,2}}$	K_α	K_α
1bkgd.°2θ	+1.00	+0.88	+1.00	+0.90	+0.88	+0.77	+1.27	+1.04
2bkgd.°2θ	-1.10	-0.80	-1.02	-1.20	-0.36	-0.27	-	-
kV	60	60	60	60	60	60	100	100
mA	32	32	32	32	32	32	20	20
Time,peak	40	40	40	40	40	40	40	40
Time,bkgd.	40	40	40	40	40	40	20	20
Bkgd.fact.	0.418	0.543	0.500	0.383	0.680	0.750	-	-

Long collimator, air path, scintillation counter, LiF (220) crystal and 3.75 window for "T". Ditto for "Mob", except gas flow proportional counter and vaccuum path.

All times in seconds.

Ba $L\beta$ was affected by Ce $L\alpha$, Zr $K\alpha$ by Sr $K\beta$, and Y $L\alpha$ by Rb $K\beta$. At high values of Cr, the Cr peak overlapped the background for Ce, so that calculated Ce concentration was diminished. No suitable correlation could be applied.

The fluorescent intensity of each selected element peak in unknown rocks was measured against that of a ratio standard, counting time or number of counts on the standard being fixed. This technique eliminates long term instrumental drift. Each determination was carried out twice. Counting time and number of counts for each element and background(s) were automatically punched in binary arithmetic on 5-hole paper tape and a visual record was kept on an arithmetic typewriter. The tapes were processed by the University of Birmingham, English Electric KDF9 computer using programs written in EGTRAN by Skinner (1970). The programs subtracted background from peak counts and divided the result by the number of counts on the ratio standard. The concentration of each element was calculated, in oxide per cent or parts per million, using the appropriate calibration coefficients and interelement correction factors for each element and block code.

Detection limits for trace elements are given below:

Cr	3	ppm
Ni	2	"
Cu	2	"
Zn	2	"
Rb	3	"
Sr	3	"
Y	3	"
Zr	3	"
Nb	3	"
Ba	6	"
La	6	"
Ce	6	"
Pb	4	"
Th	5	"

0.5 for hornblendes and peridotites; $T_{pk}, T_{bk} = 10^3$ sec.

These are limits at the 95 per cent confidence level for detection of peak above background for 100 seconds counting on the background, using relation given in Leake et al. (1969, p. 61).

For the crushed mineral fusions, calibration curves for major elements were constructed using artificial mixtures of various oxides, carbonates and chlorides, fused with Spectroflux. These covered the range of possible compositions of ferromagnesian minerals such as pyroxenes, amphiboles,

biotites, garnets, olivines and epidotes. The curves obtained were remarkably linear, and with little scatter. Calibration for trace elementswas effected by fusing spiked ultrabasic and basic rocks, whose mass absorption coefficients were assumed to approximate those of fused ferromagnesian minerals.

Precision and accuracy

The precision of the ratioing technique of X.R.F. analysis is well documented (see Leake et al., 1969, p. 68-74), all but very short term variations in operating conditions being eliminated. To check on the instrumental and replicate pellet precisions, standard samples were analysed at intervals on all machine programs. The results of these checks are summarised by Skinner (1970).

Accuracy was estimated by comparing X.R.F. analyses of various international standards, such as G1, with the accepted values. The comparisons were reasonable, except for MgO and Rb, which tended to be rather low, although good agreement was found for Rb in G1, for which the accepted value for Rb is perhaps the most reliable of the suite of international standards. SiO_2 tended to be slightly high, and Al_2O_3 was most variable.

Sources of possible error

These may be summarised as follows;

1. Particle size effect: probably the largest contribution to errors in light (Na_2O and MgO) element analysis. To some extent overcome by grinding for three minutes, but hard, foliaceous, and acicular minerals possibly remain as coarse anomalies in finer matrices.

2. Mineralogical effects are twofold. Firstly, platy and acicular minerals may be aligned on pellet surfaces, giving spurious results. The co-ordination of Al^{3+} in different minerals, such as felspar and amphibole, may affect the fluorescent yield from a sample. Calibration curves for Al_2O_3 in basic igneous and metamorphic rocks were significantly different. Apparently the percentage of hornblende in a rock controlled the fluorescent intensity of the Al K_α peak, large amounts causing anomalously low count rates.

3. Contamination was avoided by processing similar rocks in

batches, ensuring all apparatus was clean after each preparation. The apparatus was constructed of durable or "Specpure" materials. Each pellet was stored in an airtight bag, dusted before being run, and handled only by the edges. Contamination by vapourised machine oil whilst analysing, and by sea spray in the area of study has discouraged quotation of S and Cl analyses. Because of contamination discovered in Spectroflux, data for Ni and La in mineral fusions are suspect, and in the case of La, real concentration values cannot be estimated from calibration curves because of the high, unspecified concentration of La in the supposedly pure fusing compound.

References

JENKINS,R. and DE VRIES,J.L., 1967. Practical X-ray spectrometry. Phillips, Eindhoven.

LEAKE,B.E., HENDRY,G.L., KEMP,A., PLANT,A.G., HARVEY,P.K., WILSON,J.R., COATS,J.S., AUCOTT,J.W., LUNEL,T. and HOWARTH,R.J., 1969. The chemical analysis of rock powders by automatic X-ray fluorescence. Chem. Geol. 5, 7-86.

SKINNER,A.C., 1970. Unpublished Ph.D. thesis University of Birmingham.

APPENDIX 2 : MINERAL ASSEMBLAGES IN ANALYSED ROCKS

Key

QTZ	quartz	PRE	prehnite
PLA	plagioclase	ORE	opaque ore
KSP	K-felspar	RUT	rutile
OPX	orthopyroxene	ZIR	zircon
CPX	clinopyroxene	APA	apatite
1HB	primary hornblende	SPH	sphene
2HB	secondary hornblende	CCT	calcite
CUM	cummingtonite	SPI	spinel
ANT	anthophyllite	MUS	muscovite
BIO	biotite	ALB	albite
GAR	garnet	SCA	scapolite
OLI	olivine	TOU	tourmaline
SER	serpentine	ZOI	zoisite
EPI	epidote	ALL	allanite
CHL	chlorite	PUM	pumpellyite

X - present in major quantities.

O - present in minor quantities.

35 - An content of plagioclase felspar.

A - felspar completely altered and An content indeterminate.

		Q T Z	P L A	K S P	O P X	C H X	1 H B	2 N B	A U T	C M	B I O	G A R	E P I	C H L	P R E	O R E	R U T	Z I R	A P A	S P H	C C T	OTHERS
		QUARTZO-FELSPATHIC GNEISSES																				
		HYPERSTHENE-GNEISSES, TIREE																				
3020	64SW1B	X	26	0	X	X	0	0			0					0	0	0	0			
3021	65SW1A	X	24	0	0	0	0	X			X					0	0	0	0			
3022	78NE1A	X	33	0	0	0		X			X					0	0	0	0			
3023	78SW1A	X	25	0	X			0			0					0	0	0	0			
3024	78SW1B	X	26	0	X	X	X	0			0					0	0	0	0			
3025	78SW2	X	32	X	0			0								0	0	0	0			
3006	78SW5A	X	30	0	X		X	0			0					0	0	0	0			
		MIGMATITIC GNEISSES																				
2111	7344B	X	27					0			X					0	0		0	0	0	
2131	5696	X	26	0				0			X		0			0			0	0		
2133	56116E	X	29	0				X			X		0	0	0	0			0	0		
2914	70811	X	30	0							X		0	0	0	0			0	0		
2917	4007G	X	38	0				X			0		0	0	0	0			0	0		
2918	5272	X	32	0				X			X					0	0		0	0		
2919	4962A	X	28	0				0			X		0			0	0		0	0		
2920	70818C	X	25	0				0			X					0			0	0	0	
2922	56914B	X	26	0				X			X					0	0		0	0		
2923	5619	X	24	0				X			X					0			0	0		
2924	CA	X	26	0				0			X					0	0		0	0	0	
2925	CB	X	25	0				0			X					0	0		0	0		
2926	CC	X	28	0				0			X		0			0	0		0	0	0	
2927	CD	X	20	X				0			X					0	0		0	0		TOU
		NEBULITIC GNEISSES																				
2113	4931	X	31	0				0			X					0	0		0	0	0	
2130	5695	X	30	0							X					0	0		0	0		
2132	5286	X	32	0				0			X					0	0		0	0		
2152	7321F	X	31	X				0			X					0	0		0	0		
2153	7321G	X	29	X				0			X					0	0		0	0	0	
2911	4965C	X	31	X				0			X					0	0		0	0		
2912	7347	X	25	0				0			X					0	0		0	0		
2913	7066	X	31	0				X			X					0	0		0	0	0	
2915	56910A	X	30	X				X			X		0			0	0		0	0		
2916	7344A	X	21	0				0			X					0	0		0	0	0	
		DIOPSIDE-GNEISSES																				
2114	7321I	X	40	X		X		X			0					0	0		0	0		
2150	7321D	X	30	X		0		X			0					0	0			0	0	
2151	7321E	X	36	X		X		X			0					0	0		0	0		
2154	7321H	X	40	X		X		X			0					0	0			0	0	
2155	7321J	X	45	X		X		X			0					0	0		0	0		

		Q	P	K	O	C	1	2	A	C	B	G	E	C	P	O	R	Z	A	S	C	
		T	L	S	P	P	H	H	N	U	I	A	P	H	R	R	U	I	P	P	C	OTHERS
		Z	A	P	X	X	B	B	T	M	O	R	I	L	E	E	T	R	A	H	T	
		EPIDOTE-CHLORITE-GNEISSES																				
2928	5286	X	26	X				O				O		X	X	X	O			O	O	O	
2929	5551	X	30	X				O						X	X	X	O			O	O	O	MUS,ALB
2930	7341F	X	A	X										X	X	X	O			O	O	O	MUS
2931	5573A	X	A	X										X	X	X	O			O	O	O	ALB
2932	56110B	X	25	O				O						O	X	O	O			O	O	O	
2933	56110A	X	A	O										O	X	X	O			O	O	O	MUS
2934	56116A	X	A	X										X	X	O	O			O		O	
2935	56116B	X	A	X										X	X	O	O			O	O	O	ALB
2936	56116D	X	A	X				O						X	X	X	O			O	X	O	
		METASEDIMENTS ANTHOPHYLLITE-GNEISSES																				
2102	70815B	O	24					O	X			O	O				O			O			
2103	70815C	O	23						X			O					O			X	O		
2104	70815D	O	21						X			O					O			O			
2105	70815E	O	20						X			O					O			O			
		GARNET-BIOTITE-GNEISSES																				
2101	70815A	X	27					O	O			X	X				O	O					
2106	PLK	X	32					O				X	X				O			O			
2107	PLA1	X	50					O				X	X				O		O	O			TOU
2108	PLA2	X	52					O				X	X				O		O	O			TOU
2136	5288	X	32		O			X				X	O				O		O				
2156	PLGR	X	36					X				X	X				O		O	O			TOU
2157	7108	X	33									X	X				O		O	O			
2158	57810	X	26		O							X	X				O		O	O			
2159	T1D	X	47		O		O	O				X	X				O						
2850	4965A	X	52					O			X	X	X				O			O			
2867	5301S	O	47					X				X	X				O						
2998	78SE2B	X	22		X			O				O	X				O	O					
3004	78SW6A	X	32									X	X				O		O	O			
3005	64SW5A	X	35	X								X	X				O	O					SIL,KYA
		CALC-SILICATE ROCK																				
2137	56910I	O	35			O		O				X			X	X				X	O	X	SCA,ZOI
		GRANOFELSES MAFIC GRANOFELSES,TIREE																				
2991	64SW3A	O	58		X	X	X	O				O	O				O	O	O	O			
2992	64SW1A	O	47		X		X	O				O					O	O		O			
2993	64SW2A	O	50		X	X	X	O					X				O	O		O			
2994	64SW2B	O	44			X	X	O				O					O			O			
2995	78NW1B	O	32		X	O	X	O				O					O	O		O			
2996	78NW1C	O	54			X	X	O					X				O	O		O			
2997	64SW4	O	38		X	X	O	X				O					O	O		O			
2999	78SE1D	O	38		X	X	X					O	X				O	O	O	O			

		Q T Z	P L A	K S P	O P X	C P X	1 H B	2 H B	A N T	C U M	B I O	G A R	E P I	C H L	P R E	O R E	R U T	Z I R	A P A	S P H	C C T	OTHERS
3002	78NW1D	0	38			X	X	0				X			0	0		0				
3003	78SW6B	0	50		X	X	X	0			0	X			0		0	0			0	
		GARNET-BIOTITE-GRANOFELSES																				
2112	PLB	X	35								X	X		0	0	0		0	0			
2119	5724	X	33					0			X	X			0	0		0	0			
2120	5727A	X	32								X	X		0		0		0	0			
2121	5784	X	32					0			X	X			0	0		0				
2123	5301Z	X	30	0				0			X	X				0		0	0			
2124	5301Y	X	42					X			X	X						0	0			
2125	5301X	X	38			0	0	X			0	X			0							
2848	PLMA	X	45		0			0			X	X				0	0	0	0			
2849	T5A	X	42		0			0			X	X				0	0	0	0			
2855	4007E	X	40					0		0	X	X				0		0	0			
2856	4007F	X	42					X		0	X	0				0		0	0			
2857	70488	X	37					0			X	X				0		0	0			
2858	70212	0	32								X	0				0		0	0			
2869	7491	X	35					0		0	X	X				0		0	0			
3000	78SE2B	0	33	0	0						X	X				0		0	0			
3001	64NE3B	0	35	0	0						X	X				0	0	0	0			
		HORNBLENDE-BIOTITE-GRANOFELSES																				
2118	7472	X	32					0			X			0	0	0		0	0			
2122	7042	X	34								X				0	0		0	0			
2126	5301W	X	29					0		0	X					0		0	0			
2127	5301V	X	31	0		0		X			X					0		0	0	0		SCA
2128	PLE	X	32					0			X				0	0		0	0			
2129	PLD	X	A			0		X			X				0	0		0	0	0		
2134	4964A	X	35					0			X		0			0		0	0	0		
2853	70818A	X	35					X			X			0	0	0		0	0	0		
2854	B3	X	A		0			X			X		0	0	0	0		0	0	0		
2868	56914A	X	36					X			0				0	0		0	0			
2938	7042B	X	A					0			X			0	0	0		0	0			
3027	T5C	0	A			0					0			X	X			0		0		SCA
		DIOPSIDE-GRANOFELSES																				
2147	7321A	X	36	X		X		X			0			0	X	0		0	0			
2148	7321B	X	38	X		X		X			0				0	0		0	0	0		
2149	7321C	X	30	X		X		X			0				X	0			0	0		

		QTZ	PLA	KSP	OPX	CPX	1HBB	2HBB	CMU	BIO	GAR	SER	EPI	CHL	PRE	ORE	ZRU	ACT	PAS	SCH	
							METABASITES PYRIBOLITES														
2954	TIA	0	35		X	X	0	X		0						X	0		0		
2984	64NE4A	0	30		X	X	X	0		0						0	0		0		
2985	78NE2	0	56		X	X	0	X		0	X	0				0		0	0		
2986	64NE3A	0	51		X	X	0	X		0		0				0		0	0		
2987	65SW2	0	37		X	X	0	X		0		0				0			0		
2988	78SW5B		52		X	X	0	X			X	0				0	0	0	0		
2989	78NW2A	0	45		X	X	X	X		0	0					0	0	0	0		
2990	78NE1C	0	48		0	X	0	X		0		0				0			0		
3030	78NE1D	0	55		0	X	0	X		0		0				0		0	0		
							GARNET-AMPHIBOLITES														
2140	C1	0	A			X		X			0					0		0	0	0	
2141	GA	0	A			0		X			X		0			0		0	0	X	
2951	7451	0	36					X		0	X					0			0		
2952	7475	0	32			0		X		0	X					0			0		
2953	7022	0	50					X		0	X					0			0		
2955	5383	0	36			0		X			X					0			0	0	
2956	7471	0	42			0		X		0	X					0		0	0		
2957	5781A	0	42			0		X			X					0			0		
2958	4964C	0	41			X		X			0				0	0			0	0	
2961	7021O	0	44			X		X			X		0		0	0			0	0	
2963	4965D	0	A					X	0		X					0			0	0	
2964	5288	0	41			0		X	0		X					0			0		
2972	PLI		A			X		X			X				X	0			0	0	PUM
							AMPHIBOLITES														
2117	5641		A					X					0			0			0	X	
2142	GB	0	A			0		X		0						0			0	X	
2143	GC	0	35					X								0			0	0	
2851	70211A		A			0		X						0		0			0	0	
2852	7029A		A			0		X					0			0			0	0	ALB
2959	4932B	0	40	0	0			X		0		0	0			0			0	0	
2960	5374D		A					X					0	0		0			0	0	
2962	PLH	0	32			0		X		0						0			0	0	
2965	5346	0	36			0		X					0	0		0			0	X	
2966	5789	0	31			0		X								0			0	0	
2967	5592	0	35					X								0			0	0	
2968	7089		29			0		X		0			0	0		0			0		
2969	7084	0	37					X						0		0			0	0	
2970	7085		A					X								0			0	0	
2971	4007D	0	A			0		X		0			0	0		0			0		
2973	4004B	0	31					X						0		0			0	0	
2974	5781B		42			0		X								0			0		
2975	702A	0	A					X		0			0	0		0			0	0	SCA
2976	702B	0	35					X		0			0	0		0			0		
2977	5729	0	32					X								0			0	0	

		Q T Z	P L A	O L I	O P X	C P X	1 H B	2 H B	C M B	B I O	S O R	O R E	S E I	R U T	
				ULTRAMAFIC ROCKS											
2138	5552A				X	X		X		0	0	0	0	0	BIO=PHLOGOPITE
2139	5552B				X	X		X		0			0	0	
2982	56915				X	X		X		0	0	0	0	0	0 BIO=PHLOGOPITE
2983	78NW3A				X	X		X					0	0	
2979	5552C		25			X			X		0	0	0		
2980	5552D								X		X		X		2HB=TRE/ACT
2981	5552E								X		X		0		2HB=TRE/ACT

		Q T Z	P L A	K S P	O P X	G A R	B I O	M U S	O R E	Z I R	A P A	R U T	E P I	
				GRANITIC ROCKS CONCORDANT GRANITIC VEINS										
2109	C16	X	01	X						0	0			
2116	5727B	X	26	X		0					0			
2135	5289	X	24	X		0	0			0	0			
2840	PLMB	X	32	0		0	0		0	0	0			ZOI, ALL
2841	T5B	X	24	0		0	0		0		0			
2842	4965B	X	27	X		0					0			ZOI, PRE
2843	70211B	X	A	X			0			0				TOU
2844	7029B	X	A	X		0		0			0			TOU
2845	70818B	X	25	X			0		0	0	0		0	
2846	B4	X	26	0			0	0			0		0	PRE
2847	PLC	X	10	X		0	0	0	0	0	0			
2859	PLF	X	24	X			0				0			
2860	5281	X	25	X		0	0				0			HB TRACE
2861	PLJ	X	27	X			0	0	0	0	0			
2863	4007C	X	20	X				0	0		0			
2864	PLG	X	24	X			0		0		0			HB TRACE
2866	70815	X	27	0		0	0		0	0	0			
3013	64NE3C	X	22	X		0					0			
3014	64NE4B	X	22	0	0		0				0			
3015	78SE2A	X	31	X					0		0			
3016	64SW3B	X	A	X							0			
3017	78SE1C	0	30	0	0		0		0		0	0		
3018	64SW5B	X	36	X			0				0			
3019	T5D	X	15	X			0	0		0				
				REMOBILISED CONCORDANT GRANITES										
2862	S1	X	31	X					0		0			
2865	4005	X	23	X			0				0			PRE
3012	64NE1B	X	26	X					0		0			

		Q T Z	P L A	K S P X	O P R	G A R	B I O	M U S	O R E	Z I R	A P A	R U T	E P I	
							BIOTITE-PEGMATITES							
3011	64NE1A	X	30	X				0	0	0		0		
2115	7061	X	A	X				0	0			0	0	
2941	5573B	X	32	X					0			0		ALL
2942	5806	X	24	X				X		0	0	0		
2943	72210	X	22	X				X		0		0		
2945	56113	X	A	X				0		0		0		
							OTHER PEGMATITES							
2146	GG	X	36	0						0		0		
2944	7341C	X	A	X					0		0	0		
2946	4932A	X	A	X								0	0	
3010	4962B	X	20	0					0					

APPENDIX 3 : LISTING OF ROCK ANALYSES

Major and minor oxide analyses are quoted to different numbers of significant figures depending on the estimated precision and accuracy of the X.RF. method used for particular oxides. FeO, H_2O and CO_2 values are as determined by standard rapid "wet" methods.

Trace element values are given in integer form. Some are below the lower limits of detection for the machine conditions used (see p.179, Table 14, Appendix 1). Although these values are below the detection limit from theoretical calculation, they may have some significance.

C.I.P.W. norms were calculated using an EGTRAN program written by Dr. J. Tarney based on the procedure suggested by Kelsey (1965).

Tabulation was produced by an EGTRAN program written by Skinner (1970).

References.

KELSEY,C.H., 1965. Calculation of the C.I.P.W. norm. Miner. Mag. 34 (Tilley Volume), 276-282.

SKINNER,A.C., 1970. Unpublished Ph.D. thesis University of Birmingham.

HYPERSTHENE GNEISSES

ANALYSIS NUMBER -	3020	3021	3022	3023	3024	3025	3006
SAMPLE NUMBER -	64SW1B	65SW1A	78NE1A	78SW1A	78SW1B	78SW2	78SW5A
SiO2	57.8	57.9	56.0	55.5	64.1	66.8	56.0
TiO2	0.580	0.650	0.570	0.720	0.420	0.100	0.730
Al2O3	18.1	17.1	17.0	17.4	17.7	17.5	17.4
Fe2O3	4.08	3.58	3.95	3.53	1.43	0.73	3.84
FeO	2.63	3.22	3.12	3.33	1.84	0.68	2.92
MnO	0.09	0.09	0.08	0.09	0.03	0.02	0.10
MgO	2.1	2.9	2.7	3.2	1.3	0.8	2.9
CaO	5.54	5.64	5.45	6.50	3.53	1.47	6.06
Na2O	5.1	4.6	4.3	4.7	4.4	4.2	5.6
K2O	1.17	1.44	2.55	1.75	3.11	4.99	1.38
P2O5	0.310	0.321	0.271	0.377	0.172	0.095	0.390
H2O	1.06	1.00	0.99	1.03	1.01	0.70	0.91
CO2	0.51	n.d.	n.d.	0.06	0.57	0.14	0.42
TOTAL	98.7	98.6	98.0	98.2	99.5	98.2	98.6

TRACE ELEMENTS IN PARTS PER MILLION

Cr	27	24	18	39	13	27	16
Ni	18	27	17	31	8	5	15
Rb	10	20	28	20	44	74	16
Sr	1125	605	859	795	990	947	937
Yt	3	10	13	10	2	1	13
Zr	115	94	188	140	122	135	211
Nb	4	8	2	5	0	0	6
Ba	686	739	1328	867	1586	2760	499
La	27	33	47	42	17	16	44
Ce	47	75	103	93	27	18	92
Pb	12	10	12	12	14	24	23
Th	0	0	0	0	0	0	0

C.I.P.W. NORMS

Q	13.6	10.4	8.2	5.3	22.2	20.6	6.6
C	0.3	-	-	-	2.5	3.1	-
or	7.1	8.7	15.5	10.6	18.7	30.2	8.4
ab	38.0	39.8	37.3	40.1	30.5	34.8	43.2
an	25.0	22.5	20.4	22.3	16.6	6.8	21.5
nf	-	-	-	-	-	-	-
di	-	3.3	4.5	6.6	-	-	5.3
hy	5.9	7.0	6.4	7.2	4.7	2.6	5.9
ol	-	-	-	-	-	-	-
mt	6.0	5.3	5.9	5.3	2.1	1.1	5.7
il	1.1	1.3	1.1	1.4	0.8	0.2	1.4
hm	-	-	-	-	-	-	-
ap	0.8	0.8	0.7	0.9	0.4	0.2	0.0

MIGMATITIC GNEISSES

ANALYSIS NUMBER -	2111	2131	2133	2914	2917	2918	2919	2920	2922
SAMPLE NUMBER -	7344B	5696	56116E	73811	4007G	50272	4962A	70818C	56914B
SiO2	60.6	64.6	62.4	53.9	62.8	66.2	62.9	61.7	62.1
TiO2	0.630	0.540	0.520	0.600	0.570	0.410	0.540	0.610	0.550
Al2O3	16.5	15.1	15.4	18.5	15.2	15.4	15.6	14.5	14.6
Fe2O3	3.87	2.77	3.46	2.87	2.98	3.34	2.19	3.41	3.44
FeO	2.03	2.80	2.04	2.90	2.98	0.86	3.09	3.12	2.82
MnO	0.07	0.06	0.06	0.05	0.07	0.05	0.07	0.06	0.08
MgO	2.5	2.3	3.1	3.5	2.3	2.1	2.3	2.0	2.9
CaO	4.30	2.43	3.74	3.43	4.87	3.27	3.59	3.49	3.90
Na2O	3.8	3.0	3.6	4.8	3.4	3.9	3.9	3.5	3.4
K2O	1.99	2.88	2.14	2.86	1.69	2.00	2.07	2.11	2.58
P2O5	0.151	0.155	0.092	0.033	0.100	0.085	0.121	0.095	0.081
H2O	1.39	1.16	0.76	1.08	1.71	1.07	2.54	1.70	2.68
CO2	0.52	0.24	0.26	0.78	0.57	0.71	0.60	0.71	0.46
TOTAL	98.4	98.9	97.6	95.3	99.2	99.5	99.5	98.0	99.5

TRACE ELEMENTS IN PARTS PER MILLION

Cr	71	107	139	52	56	99	97	154	169
Ni	47	55	93	37	38	57	47	83	86
Rb	92	129	102	123	70	86	103	96	94
Sr	377	278	394	437	208	397	364	311	286
Yt	7	5	5	0	13	6	5	2	11
Zr	165	233	134	290	161	144	134	175	135
Nb	4	7	6	9	12	6	5	10	5
Ba	436	573	491	574	246	622	344	755	597
La	38	19	26	1	18	22	17	31	17
Ce	76	33	42	2	34	39	46	52	28
Pb	12	9	17	18	12	13	12	10	12
Th	11	10	6	4	11	4	6	18	4

C.I.P.W. NORMS

Q	23.5	23.9	23.7	8.5	27.2	32.3	25.8	27.3	23.0
C	1.9	2.0	1.2	3.3	0.5	2.8	2.1	2.0	0.4
or	12.1	17.4	13.1	17.9	10.2	12.0	12.6	12.0	15.7
ab	26.7	31.3	28.4	33.2	22.7	25.1	26.8	22.2	23.9
an	20.0	11.3	18.5	17.8	24.1	15.9	17.6	17.3	19.4
nf	-	-	-	-	-	-	-	-	-
di	-	-	-	-	-	-	-	-	-
hy	6.3	7.9	8.0	11.4	8.0	5.2	8.0	9.8	8.0
ol	-	-	-	-	-	-	-	-	-
mt	5.1	4.1	5.2	4.4	4.4	1.7	3.3	5.1	5.1
il	1.2	1.0	1.0	1.2	1.1	0.8	1.1	1.2	1.1
hm	0.5	-	-	-	-	2.2	-	-	-
ap	0.4	0.4	0.2	0.1	0.2	0.2	0.3	0.2	0.2

5 MIGMATITIC 4 NEBULITIC GNEISSES

ANALYSIS NUMBER -	2923	2924	2925	2926	2927	2113	2130	2132	2152
SAMPLE NUMBER -	5619	CA	CB	CC	CD	4931	5695	5286	7321F
SiO2	60.5	63.7	62.9	65.8	67.1	66.0	68.6	66.4	67.5
TiO2	0.630	0.560	0.580	0.450	0.350	0.290	0.250	0.380	0.180
Al203	15.2	14.8	14.6	14.9	16.5	16.6	17.1	16.5	16.5
Fe203	3.55	3.45	2.92	1.74	0.27	1.86	0.87	1.35	1.66
FeO	3.15	2.18	2.98	2.89	2.96	1.29	1.13	1.90	0.84
MnO	0.09	0.07	0.07	0.05	0.04	0.04	0.02	0.03	0.04
MgO	2.8	3.1	3.1	2.1	1.1	1.0	0.9	1.1	1.0
CaO	4.26	2.39	2.43	2.64	3.24	3.78	3.41	3.61	1.75
Na2O	3.5	3.9	3.7	3.7	4.5	4.4	4.7	4.8	4.0
K2O	2.29	2.36	2.72	2.67	2.12	2.53	1.79	1.46	4.99
P2O5	0.146	0.126	0.112	0.126	0.112	0.103	0.063	0.104	0.096
H2O	1.78	1.30	1.71	1.59	0.97	0.60	0.63	0.97	0.84
CO2	0.18	0.38	0.28	0.49	0.38	0.67	0.25	0.38	0.19
TOTAL	98.1	98.4	98.1	99.2	99.5	100.1	99.7	98.0	99.0

TRACE ELEMENTS IN PARTS PER MILLION

Cr	57	110	126	98	21	11	32	19	15
Ni	32	52	65	56	18	15	17	13	11
Rb	90	118	128	113	78	78	60	52	116
Sr	320	267	282	336	344	393	423	439	467
Yt	13	8	8	3	8	7	0	6	10
Zr	95	175	149	146	171	234	181	211	134
Nb	10	7	12	7	5	6	5	3	0
Ba	471	463	579	702	647	659	529	361	1568
La	27	22	28	24	32	38	15	40	49
Ce	51	35	47	44	45	64	29	74	99
Pb	12	10	20	9	13	26	17	17	14
Th	6	13	14	9	11	9	12	7	13

C.I.P.W. NORMS

Q	19.9	26.1	23.6	28.7	25.9	27.0	27.0	26.5	21.8
C	0.1	2.7	2.1	2.6	2.1	1.6	1.0	1.6	2.0
or	14.0	14.4	16.7	16.2	12.7	15.2	10.7	8.8	29.8
ab	28.2	29.8	29.2	26.4	33.9	29.7	36.0	36.9	32.1
an	20.0	11.4	11.8	12.6	15.5	18.3	16.7	17.6	8.1
nf	-	-	-	-	-	-	-	-	-
di	-	-	-	-	-	-	-	-	-
hy	9.4	8.2	10.2	8.6	7.6	3.0	3.3	4.6	2.7
ol	-	-	-	-	-	-	-	-	-
mt	5.4	5.1	4.4	2.6	0.4	2.7	1.3	2.0	2.3
il	1.2	1.1	1.1	0.9	0.7	0.6	0.5	0.7	0.4
hm		-	-	-	-	-	-	-	0.1
ap	0.4	0.3	0.3	0.3	0.3	0.3	0.2	0.3	0.2

NEBULITIC GNEISSES

ANALYSIS NUMBER -	2153	2911	2912	2913	2915	2916	2921	3026
SAMPLE NUMBER -	7321G	4965C	7347	7066	56910A	7344A	4001	65SW1B
SiO2	68.7	66.6	69.8	68.6	69.2	65.8	68.4	63.7
TiO2	0.250	0.390	0.220	0.280	0.180	0.370	0.370	0.580
Al2O3	16.2	16.5	15.9	16.6	16.0	16.2	16.2	15.4
Fe2O3	2.28	1.96	1.04	1.78	0.21	1.92	1.51	3.29
FeO	0.50	1.12	1.32	1.06	1.35	2.18	1.49	1.74
MnO	0.04	0.04	0.04	0.04	0.02	0.06	0.04	0.05
MgO	1.0	1.1	0.8	0.8	0.5	1.4	1.0	1.8
CaO	2.94	3.50	3.06	2.86	3.26	4.29	3.70	3.58
Na2O	4.9	4.6	4.7	4.8	4.7	4.4	4.1	4.2
K2O	1.81	2.09	1.99	1.98	1.73	1.58	2.05	2.82
P2O5	0.124	0.105	0.073	0.056	0.022	0.091	0.077	0.177
H2O	n.d.	0.51	1.16	0.09	0.73	0.33	0.88	1.30
CO2	0.20	0.26	2.07	0.12	0.46	0.23	0.46	0.26
TOTAL	99.0	98.8	102.1	99.1	99.4	98.8	100.3	99.8

TRACE ELEMENTS IN PARTS PER MILLION

Cr	14	51	18	17	14	35	21	20
Ni	20	22	9	9	8	30	15	11
Rb	63	77	53	87	47	51	67	50
Sr	368	520	600	341	490	329	367	619
Yt	5	8	6	5	3	9	4	15
Zr	170	168	166	174	147	156	148	240
Nb	5	8	3	3	0	4	6	6
Ba	649	583	809	559	663	431	726	1125
La	62	33	33	26	13	13	20	35
Ce	117	45	68	50	45	40	22	75
Pb	17	14	6	15	30	15	17	16
Th	22	3	20	20	12	8	8	2

C.I.P.W. NORMS

Q	26.8	25.9	43.3	26.9	30.9	24.6	30.9	21.4
C	1.5	1.3	5.4	1.8	2.6	0.2	1.8	-
or	10.7	12.6	11.6	11.8	10.4	9.5	12.2	17.1
ab	39.4	36.1	15.0	39.9	34.8	34.9	29.4	33.1
an	13.8	16.0	14.5	13.0	16.3	21.0	17.0	16.9
nf	-	-	-	-	-	-	-	-
di	-	-	-	-	-	-	-	0.1
hy	2.5	2.7	3.1	2.0	3.3	5.5	3.6	4.5
ol	-	-	-	-	-	-	-	-
mt	1.0	2.7	1.5	2.6	0.3	2.8	2.2	4.2
il	0.5	0.8	0.4	0.5	0.4	0.7	0.7	1.1
hm	1.6	0.2	-	-	-	-	-	0.5
ap	0.3	0.3	0.2	0.1	0.1	0.2	0.2	0.4

PYRIBOLITES

ANALYSIS NUMBER -	2954	2984	2985	2986	2987	2988	2989	2990	3030
SAMPLE NUMBER -	T1A	64NE4A	78NE2	64NE3A	65SW2	78SW5B	78NW2A	78NE1C	78NE1D
SiO2	50.9	49.5	48.6	50.4	51.3	50.7	48.6	48.3	48.6
TiO2	2.801	1.740	1.089	1.286	2.456	1.012	1.063	1.145	1.165
Al2O3	14.2	11.9	15.2	14.6	15.9	16.8	16.5	14.9	14.8
Fe2O3	3.75	6.02	5.38	5.74	9.33	4.94	3.92	6.00	5.84
FeO	8.94	8.91	9.20	6.34	7.26	8.21	7.61	6.67	6.98
MnO	0.21	0.20	0.20	0.13	0.17	0.18	0.23	0.17	0.17
MgO	6.1	6.3	7.4	6.7	5.4	6.4	7.4	8.1	7.9
CaO	3.63	9.64	6.24	5.17	3.87	6.25	5.63	5.94	5.79
Na2O	3.2	3.5	3.0	4.0	3.9	2.5	3.2	3.1	3.1
K2O	0.41	0.59	0.23	0.87	0.78	0.33	0.76	0.59	0.85
P2O5	0.174	0.112	0.069	0.188	0.165	0.084	0.061	0.071	0.070
H2O	0.79	1.88	2.07	3.51	1.54	0.89	3.47	3.24	3.20
CO2	0.32	0.73	0.59	0.72	0.56	0.58	0.39	0.27	0.70
TOTAL	100.5	101.1	100.3	99.7	102.7	99.9	98.8	98.5	99.2

TRACE ELEMENTS IN PARTS PER MILLION

Cr	120	121	282	309	193	161	404	258	223
Ni	97	89	99	94	91	68	83	118	107
Cu	215	136	148	49	282	108	34	108	70
Zn	82	133	82	28	118	75	93	76	86
Rb	6	6	6	11	7	6	14	7	6
Sr	108	166	99	508	254	162	216	104	321
Yt	37	22	19	12	27	18	19	16	17
Zr	147	80	55	93	173	56	66	62	67
Nb	10	4	2	3	10	4	1	3	5
Ba	181	225	112	209	403	180	239	100	224
La	20	8	0	21	21	5	6	2	8
Ce	42	4	0	33	48	11	12	14	17
Pb	9	12	12	11	4	13	8	4	11
Th	3	2	2	3	0	0	0	1	1

C.I.P.W. NORMS

Q	15.0	4.9	1.1	7.3	13.1	12.9	3.5	3.6	6.3
C	3.2	-	-	-	3.2	2.6	1.3	-	0.0
or	2.4	3.5	1.4	5.4	4.6	1.0	4.7	3.7	5.2
ab	22.9	21.3	27.1	26.3	26.4	14.3	23.2	24.3	18.9
an	16.9	19.7	27.2	24.8	17.9	31.1	28.9	27.0	29.4
nf	-	-	-	-	-	-	-	-	-
di	-	22.6	3.2	0.5	-	-	-	1.0	-
hy	20.3	13.9	28.4	22.4	15.1	26.2	29.2	26.2	27.0
ol	-	-	-	-	-	-	-	-	-
mt	12.7	8.8	7.9	8.7	13.4	7.3	5.0	9.1	8.8
il	5.3	3.3	2.1	2.5	4.6	1.0	2.1	2.3	2.3
hm	-	-	-	-	-	-	-	-	-
ap	0.4	0.3	0.2	0.5	0.4	0.2	0.2	0.2	0.2

GARNET AMPHIBOLITES

ANALYSIS NUMBER -	2140	2141	2951	2952	2953	2955	2956	2957	2958
SAMPLE NUMBER -	C1	GA	7451	7475	7022	5383	7471	5781A	4964C
SiO2	48.7	48.4	53.9	52.3	48.3	48.6	52.5	49.7	47.8
TiO2	1.160	0.660	1.067	1.043	1.320	1.115	2.533	1.427	0.846
Al2O3	12.5	14.0	10.8	10.8	14.8	13.7	12.6	14.5	15.6
Fe2O3	9.32	2.39	7.56	6.89	4.23	5.71	4.85	6.03	5.74
FeO	5.10	7.70	13.27	13.74	9.96	8.97	12.00	8.34	8.65
MnO	0.22	0.28	0.22	0.22	0.26	0.18	0.25	0.18	0.52
MgO	5.0	4.4	5.0	5.4	7.1	8.8	5.7	6.5	5.6
CaO	10.04	13.49	4.02	4.11	5.39	5.46	4.11	5.77	6.33
Na2O	2.3	2.2	1.2	1.2	2.5	1.8	1.2	2.1	2.4
K2O	1.33	2.27	0.64	0.68	0.70	0.69	1.43	0.56	1.10
P2O5	0.090	0.107	0.041	0.043	0.081	0.060	0.260	0.084	0.048
H2O	3.71	2.37	2.16	2.77	2.71	3.51	2.14	3.15	2.06
CO2	0.39	0.72	0.33	0.33	0.14	0.51	0.11	0.54	1.74
TOTAL	100.8	100.0	100.8	100.1	97.4	99.1	99.5	98.9	98.4

TRACE ELEMENTS IN PARTS PER MILLION

Cr	289	243	23	21	190	267	88	183	404
Ni	149	138	19	26	96	131	64	102	177
	106	32	14	3	55	54	9	37	0
	87	107	33	34	91	85	107	164	0
Rb	58	104	17	12	21	9	29	13	13
Sr	183	223	33	32	156	81	80	193	225
Yt	23	20	17	13	22	22	35	19	15
Zr	66	58	46	44	78	63	167	80	40
Nb	2	2	0	2	4	1	14	3	3
Ba	366	478	170	183	149	116	385	161	161
	0	10	7	0	3	22	29	2	0
Ce	0	30	0	3	8	7	57	8	0
Pb	9	23	6	8	18	4	7	14	25
Th	1	0	1	0	4	0	6	0	2

C.I.P.W. NORMS

Q	9.3	0.5	28.1	21.9	4.5	10.9	18.2	14.0	17.0
C	-	-	2.9	1.7	0.6	1.5	2.6	1.5	3.2
or	8.1	13.7	3.8	4.1	4.4	4.3	8.7	3.5	6.7
ab	15.2	10.1	0.3	5.8	20.7	9.5	9.1	11.6	-
an	23.0	29.6	19.0	20.7	27.6	27.9	19.2	29.3	32.3
nf	-	-	-	-	-	-	-	-	-
di	21.6	31.7	-	-	-	-	-	-	-
hy	5.5	7.5	29.6	32.4	32.5	33.6	29.3	25.6	25.7
ol	-	-	-	-	-	-	-	-	-
mt	13.9	3.5	11.1	10.3	6.5	8.7	7.2	9.1	8.6
il	2.3	1.3	2.1	2.0	2.6	2.2	4.9	2.8	1.7
hm	-	-	-	-	-	-	-	-	-
ap	0.2	0.3	0.1	0.1	0.2	0.2	0.6	0.2	0.1

4 GARNET AMPHIBOLITES; 5 AMPHIBOLITES

ANALYSIS NUMBER -	2961	2963	2964	2972	2117	2142	2143	2851	2852
SAMPLE NUMBER -	70210	4965D	5288	PLI	5641	GB	GC	70211A	7029A
SiO2	50.7	51.1	50.4	43.3	48.9	49.7	53.1	56.2	48.9
TiO2	1.285	1.804	2.425	1.068	0.900	0.380	0.140	0.780	0.600
Al2O3	18.4	13.0	12.8	17.6	12.2	14.3	5.3	14.4	13.1
Fe2O3	3.67	5.35	6.89	7.00	3.68	1.56	3.57	3.28	4.02
FeO	6.04	10.92	10.27	8.44	7.10	7.93	7.93	4.69	5.99
MnO	0.21	0.21	0.25	0.21	0.21	0.18	0.37	0.14	0.19
MgO	3.9	7.4	5.9	8.2	5.3	8.6	14.2	6.2	5.2
CaO	6.30	4.35	4.42	5.98	16.17	8.79	12.59	8.34	12.34
Na2O	3.3	1.1	1.5	1.1	1.9	3.1	0.8	2.7	2.8
K2O	0.76	1.00	1.54	2.22	1.02	1.82	0.41	1.28	1.38
P2O5	0.097	0.164	0.223	0.082	0.079	0.032	0.002	0.165	0.050
H2O	2.28	1.93	2.44	2.47	2.10	2.41	1.42	1.32	3.17
CO2	0.53	0.49	1.53	1.57	1.11	0.52	0.58	0.17	0.39
TOTAL	98.4	98.9	99.0	99.3	100.6	99.3	100.4	99.7	98.1

TRACE ELEMENTS IN PARTS PER MILLION

Cr	206	229	100	442	214	167	1237	126	225
Ni	97	106	62	137	121	153	976	82	120
Cu	44	45	66	183	58	21	0	0	0
Zn	116	132	119	107	76	114	163	0	0
Rb	29	32	35	82	51	72	3	38	41
Sr	287	23	99	91	196	407	13	185	135
Yt	26	23	41	22	21	17	14	22	11
Zr	95	104	186	71	62	37	19	173	36
Nb	5	7	12	2	2	2	5	6	5
Ba	220	325	348	259	172	382	74	248	1023
La	4	10	26	2	2	0	0	14	1
Ce	12	27	64	2	13	17	11	38	7
Pb	27	19	24	19	15	25	1	26	27
Th	3	2	0	0	6	0	2	0	3

C.I.P.W. NORMS

Q	11.6	19.5	26.7	13.1	7.8	-	8.7	11.7	1.0
C	2.3	3.8	4.7	6.6	-	-	-	-	-
or	4.7	6.1	9.3	13.5	6.1	11.1	2.4	7.7	8.6
ab	22.7	3.8	-	-	2.8	20.6	0.0	21.2	19.8
an	32.2	21.2	20.9	30.1	29.2	23.9	13.5	24.8	22.0
nf	-	-	-	-	-	-	-	-	-
di	-	-	-	-	42.3	17.0	39.8	12.9	33.4
hy	16.8	32.6	24.7	29.7	1.7	11.8	28.7	14.5	4.9
ol	-	-	-	-	-	11.2	-	-	-
mt	5.6	8.0	10.2	10.5	5.4	2.3	5.2	4.8	6.1
il	2.6	3.5	4.7	2.1	1.7	0.7	0.3	1.5	1.2
hm	-	-	-	-	-	-	-	-	-
ap	0.2	0.4	0.5	0.2	0.2	0.1	-	0.4	0.1

AMPHIBOLITES

ANALYSIS NUMBER -	2959	2960	2962	2965	2966	2967	2968	2969	2970
SAMPLE NUMBER -	4932B	5374	PLH	5346	5789	5592	7089	7084	7085
SiO2	48.7	48.7	50.9	48.7	50.2	49.5	49.6	50.8	50.2
TiO2	2.257	1.239	0.972	1.569	1.386	1.954	2.146	1.173	1.098
Al2O3	12.6	12.5	15.1	13.1	14.2	13.9	12.5	13.4	13.0
Fe2O3	7.75	7.84	4.98	6.68	8.53	7.31	7.29	7.48	6.59
FeO	9.59	9.54	7.18	8.57	7.87	9.26	10.04	8.09	8.83
MnO	0.21	0.13	0.20	0.19	0.20	0.20	0.29	0.23	0.22
MgO	5.3	6.3	7.1	5.8	5.7	5.5	4.0	5.7	6.1
CaO	4.65	4.75	5.28	6.28	5.12	4.86	4.59	4.54	5.05
Na2O	2.3	2.9	2.9	2.6	3.3	2.3	2.3	2.5	2.3
K2O	1.58	1.12	0.95	0.36	0.81	1.23	1.32	1.33	1.23
P2O5	0.232	0.059	0.069	0.141	0.118	0.204	0.181	0.082	0.069
H2O	2.03	3.01	1.28	2.97	0.72	2.65	2.46	2.03	2.73
CO2	0.66	0.91	1.28	0.72	0.50	1.50	1.87	0.73	0.50
TOTAL	97.7	98.9	98.1	97.7	99.2	100.5	99.6	98.0	98.5

TRACE ELEMENTS IN PARTS PER MILLION

Cr	101	74	204	93	86	116	74	56	74
Ni	61	46	78	52	64	60	47	57	52
Cu	0	136	0	284	35	350	107	37	30
Zn	0	15	116	79	96	116	116	108	92
Rb	32	11	22	6	20	40	22	27	16
Sr	140	78	112	158	152	260	265	184	150
Yt	34	22	23	27	26	31	33	19	19
Zr	157	45	74	87	104	160	137	82	65
Nb	11	55	5	6	4	11	9	6	3
Ba	361	174	157	154	247	405	320	172	182
La	31	0	11	2	14	22	21	17	6
Ce	51	0	23	5	24	39	33	36	9
Pb	15	12	40	18	22	20	23	11	8
Th	1	0	4	1	7	0	4	3	0

C.I.P.W. NORMS

Q	14.7	11.8	16.4	13.9	10.5	22.2	25.9	16.8	12.8
C	0.8	0.1	2.0	-	0.0	3.0	3.7	1.6	0.0
or	9.7	6.9	5.8	2.2	4.9	7.4	8.0	8.2	7.6
ab	11.7	14.4	9.2	14.3	22.5	2.1	-	12.6	14.4
an	22.5	24.2	26.6	29.0	25.1	23.3	22.2	22.9	25.8
nf	-	-	-	-	-	-	-	-	-
di	-	-	-	2.3	-	-	-	-	-
hy	22.1	25.9	26.3	22.6	20.1	22.2	22.5	22.3	25.7
ol	-	-	-	-	-	-	-	-	-
mt	11.7	11.9	7.4	10.2	12.6	10.8	10.9	11.3	10.0
il	4.5	2.4	1.9	3.1	2.7	3.8	4.2	2.3	2.2
hm	-	-	-	-	-	-	-	-	-
ap	0.6	0.2	0.2	0.4	0.3	0.5	0.4	0.2	0.2

AMPHIBOLITES

ANALYSIS NUMBER -	2971	2973	2974	2975	2976	2977
SAMPLE NUMBER -	4007D	4004B	5781B	7029A	7029B	5729
SiO2	52.5	48.5	49.3	47.7	48.7	49.7
TiO2	1.834	0.552	1.520	1.876	1.847	2.673
Al2O3	14.7	11.1	14.1	13.4	13.3	12.3
Fe2O3	6.99	5.76	5.65	6.59	8.12	7.00
FeO	8.44	6.64	8.50	8.59	7.82	10.41
MnO	0.22	0.21	0.19	0.22	0.21	0.21
MgO	5.1	11.8	6.4	6.9	6.5	4.7
CaO	3.98	6.23	5.62	4.94	5.30	4.83
Na2O	2.3	1.0	2.1	2.3	2.7	2.2
K2O	1.22	0.83	0.67	1.37	1.16	1.19
P2O5	0.185	0.035	0.095	0.156	0.161	0.282
H2O	2.02	4.70	3.23	3.57	3.41	4.32
CO2	0.68	0.32	0.17	0.43	0.57	0.71
TOTAL	100.2	98.6	97.4	98.1	99.4	100.4

TRACE ELEMENTS IN PARTS PER MILLION

Cr	90	1014	188	192	188	89
Ni	63	253	111	74	72	50
Cu	41	43	98	275	92	149
Zn	106	79	111	109	101	118
Rb	23	7	11	50	18	14
Sr	135	112	139	186	173	153
Yt	25	13	20	26	23	36
Zr	145	38	88	117	118	182
Nb	11	2	5	12	12	14
Ba	246	131	230	313	332	394
La	19	0	5	9	11	33
Ce	39	0	4	32	21	76
Pb	29	16	7	9	13	11
Th	4	0	1	0	12	2

C.I.P.W. NORMS

Q	21.4	4.9	11.3	9.6	11.2	18.3
C	4.4	-	0.3	0.7	-	0.0
or	7.3	5.2	4.2	8.6	7.1	7.3
ab	11.8	13.9	16.7	14.8	16.3	10.3
an	18.9	22.2	28.9	24.9	25.5	23.0
nf	-	-	-	-	-	-
di	-	8.3	-	-	0.6	-
hy	20.2	34.6	26.2	26.2	21.7	21.8
ol	-	-	-	-	-	-
mt	10.3	8.9	8.7	10.1	12.2	10.6
il	3.6	1.1	3.1	3.8	3.6	5.3
hm	-	-	-	-	-	-
ap	0.5	0.1	0.2	0.4	0.4	0.7

ULTRAMAFIC ROCKS

ANALYSIS NUMBER -	2138	2139	2982	2983	2979	2980	2981
SAMPLE NUMBER -	5552A	5552B	56915	78NW3B	5552C	5552D	5552E
SiO2	47.1	46.1	41.4	41.9	48.3	53.7	49.6
TiO2	0.240	0.220	0.400	1.060	0.770	0.170	0.260
Al2O3	5.9	5.2	6.5	3.4	9.7	5.5	7.0
Fe2O3	3.96	4.06	6.08	5.93	2.76	2.60	2.67
FeO	5.89	6.01	9.53	7.10	8.16	5.66	7.04
MnO	0.20	0.19	0.23	0.22	0.19	0.20	0.28
MgO	24.3	24.6	25.5	25.2	14.5	19.6	21.7
CaO	7.18	5.06	3.56	6.57	8.17	11.90	4.42
Na2O	0.5	0.5	0.8	0.6	1.8	0.7	0.4
K2O	0.108	0.092	0.612	0.291	0.496	0.382	2.580
P2O5	0.010	0.006	0.075	0.065	0.068	0.016	0.007
H2O	2.62	4.15	3.12	3.00	3.18	3.16	3.10
CO2	1.89	1.55	1.67	1.00	1.21	1.00	0.81
TOTAL	99.9	97.7	99.6	96.4	99.4	104.6	100.2

TRACE ELEMENTS IN PARTS PER MILLION

Cr	2233	2191	77	1685	1018	3040	2167
Ni	1749	1905	338	1706	662	1733	1715
Rb	3	1	21	3	5	27	180
Sr	42	37	108	56	42	19	11
Yt	6	7	5	9	8	3	14
Zr	15	16	41	59	58	17	19
Nb	1	1	2	3	0	0	4
Ba	124	65	192	107	74	68	189
La	0	0	7	11	1	1	0
Ce	0	0	0	0	0	0	0
Pb	10	17	18	12	5	4	1
Th	0	4	0	1	0	0	0

C.I.P.W. NORMS

Q	6.5	5.0	-	-	4.8	6.6	-
C	-	-	2.3	-	-	-	-
or	0.7	0.6	3.8	1.8	3.1	2.2	15.7
ab	-	-	-	-	1.2	-	-
an	26.1	23.2	17.8	12.9	25.4	16.7	15.2
nf	-	-	-	-	-	-	-
di	8.2	2.8	-	16.0	13.0	32.8	5.9
hy	65.0	72.3	68.0	49.1	43.7	40.8	59.9
ol	-	-	6.9	12.4	-	-	3.1
mt	5.9	6.2	9.1	9.2	4.2	3.7	4.0
il	0.5	0.5	0.8	2.2	1.5	0.3	0.5
hm	-	-	-	-	-	-	-
ap	0.0	0.0	0.2	0.2	0.2	0.0	0.0

CONCORDANT GRANITIC VEINS

ANALYSIS NUMBER -	2109	2116	2135	2840	2841	2842	2843	2844	2845
SAMPLE NUMBER -	C16	5727B	5289	PLMB	T5B	4965B	70211B	7029B	70818B
SiO2	67.7	71.7	74.3	78.6	77.8	73.8	74.5	76.1	75.9
TiO2	0.260	0.012	0.070	0.020	0.099	0.033	0.005	0.046	0.040
Al203	14.1	17.1	15.5	14.2	13.6	16.3	16.3	14.8	15.3
Fe203	1.39	0.00	0.17	0.19	0.84	0.61	0.00	0.35	0.53
FeO	0.73	0.68	0.66	0.63	0.78	0.52	0.27	0.54	0.14
MnO	0.03	0.06	0.02	0.03	0.05	0.04	0.02	0.02	0.01
MgO	3.2	0.0	0.0	0.1	0.5	0.2	0.0	0.2	0.1
CaO	5.85	1.42	1.02	3.31	2.59	4.01	0.80	1.16	3.13
Na20	3.7	3.5	4.2	4.4	4.4	4.8	2.5	2.2	4.2
K20	1.58	6.01	4.59	0.51	0.83	0.51	7.85	6.50	1.27
P205	0.076	0.013	0.008	0.012	0.013	0.014	0.032	0.022	0.000
H20	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CO2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
TOTAL	98.7	100.5	100.6	102.0	101.4	100.9	102.3	102.4	100.6

TRACE ELEMENTS IN PARTS PER MILLION

Cr	32	7	7	15	12	17	7	5	13
Ni	18	6	1	6	5	8	0	4	15
Cu	0	0	8	25	30	17	26	11	28
Zn	0	0	19	15	53	0	2	11	2
Rb	28	108	201	9	30	14	335	161	32
Sr	152	260	79	285	180	433	71	112	468
Yt	3	34	34	14	19	5	13	9	1
Zr	140	98	79	198	130	81	5	119	120
Nb	1	2	11	1	4	2	1	3	4
Ba	596	923	146	153	141	229	182	219	521
La	11	1	1	3	8	5	0	34	2
Ce	25	3	8	3	8	8	0	86	5
Pb	12	73	35	71	75	48	47	88	21
Th	2	0	17	3	5	2	5	39	3

C.I.P.W. NORMS

Q	25.6	24.8	29.3	42.2	41.0	34.6	27.3	34.6	39.5
C	-	2.3	1.8	0.3	0.8	0.6	2.3	2.0	1.3
or	9.5	35.3	26.0	2.9	4.8	2.0	45.3	37.7	7.5
ab	31.0	29.1	35.6	36.9	36.5	40.2	20.8	18.4	35.3
an	17.4	6.9	4.0	16.0	12.6	19.6	3.7	5.5	15.4
nf	-	-	-	-	-	-	-	-	-
di	8.0	-	-	-	-	-	-	-	-
hy	3.0	1.4	1.0	1.3	1.9	0.0	0.6	1.2	0.3
ol	-	-	-	-	-	-	-	-	-
mt	1.7	-	0.3	0.3	1.2	0.9	-	0.5	0.4
il	0.5	0.0	0.1	0.0	0.2	0.1	0.0	0.1	0.1
hm	0.2	-	-	-	-	-	-	-	0.3
ap	0.2	0.0	0.0	0.0	0.0	0.0	0.1	0.1	-

CONCORDANT GRANITIC VEINS

ANALYSIS NUMBER -	2846	2847	2859	2860	2861	2863	2864	2866	3013
SAMPLE NUMBER -	B4	PLC	PLF	5281	PLJ	4007C	PLG	70815	64NE3C
SiO2	73.0	75.3	72.6	73.6	73.2	72.7	73.4	76.6	75.9
TiO2	0.093	0.022	0.123	0.068	0.020	0.082	0.076	0.062	0.027
Al2O3	15.9	15.3	15.4	15.5	16.6	15.5	15.1	13.7	14.5
Fe2O3	0.44	0.19	1.00	0.15	0.15	0.35	0.71	0.47	0.18
FeO	0.46	0.52	0.54	0.71	0.46	0.60	0.40	0.40	0.33
MnO	0.01	0.03	0.02	0.03	0.01	0.01	0.03	0.01	0.01
MgO	0.3	0.1	0.3	0.0	0.2	0.6	0.3	0.2	0.0
CaO	3.35	0.73	1.48	0.99	4.45	3.00	2.62	2.64	0.82
Na2O	4.0	3.2	3.6	4.2	4.7	4.7	5.3	4.9	5.1
K2O	0.93	5.90	4.95	4.71	0.55	1.15	1.42	0.66	3.09
P2O5	0.008	0.020	0.119	0.005	0.013	0.004	0.050	0.020	0.008
H2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CO2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
TOTAL	100.4	101.2	100.2	99.0	100.4	98.6	99.4	99.6	100.0

TRACE ELEMENTS IN PARTS PER MILLION

Cr	11	73	8	7	23	19	28	10	9
Ni	5	4	3	0	11	23	5	7	0
Cu	11	15	35	14	58	10	2200	178	17
Zn	16	19	21	9	0	1	9	0	0
Rb	42	156	184	190	13	38	46	23	67
Sr	291	121	145	83	170	376	189	119	82
Yt	0	13	9	12	5	2	12	1	3
Zr	161	139	159	55	71	254	53	24	90
Nb	2	2	8	7	1	3	6	3	0
Ba	155	475	534	107	115	243	250	177	286
La	7	13	15	1	4	2	2	8	1
Ce	1	28	30	12	6	3	98	6	8
Pb	21	141	38	47	29	21	30	180	45
Th	6	12	12	19	0	5	17	10	4

C.I.P.W. NORMS

Q	33.5	31.6	29.4	28.6	33.5	33.0	31.6	39.9	32.7
C	0.6	2.3	1.7	1.7	0.2	1.1	0.2	0.2	1.4
or	5.5	34.4	29.2	27.8	3.2	6.9	8.5	3.9	18.3
ab	41.9	26.9	30.4	35.4	39.7	40.2	44.0	41.3	42.8
an	16.5	3.4	6.6	4.9	21.9	15.1	12.8	13.0	4.0
nf	-	-	-	-	-	-	-	-	-
di	-	-	-	-	-	-	-	-	-
hy	1.2	0.0	0.8	1.2	1.2	2.2	0.7	0.8	0.5
ol	-	-	-	-	-	-	-	-	-
mt	0.6	0.3	1.5	0.2	0.2	0.5	1.0	0.7	0.3
il	0.2	0.0	0.2	0.1	0.0	0.2	0.2	0.1	0.1
hm	-	-	-	-	-	-	-	-	-
ap	0.0	0.1	0.3	0.0	0.0	0.0	0.1	0.1	0.0

6 CONCORDANT; 3 REMOBILISED GRANITIC VEINS

ANALYSIS NUMBER -	3014	3015	3016	3017	3018	3019	2862	2865	3012
SAMPLE NUMBER -	64NE4B	78SE2A	64SW3B	78SE1C	64SW5B	T5D	S1	4005	64NE1B
SiO2	73.1	74.8	76.1	56.4	69.0	74.4	76.2	73.3	74.3
TiO2	0.207	0.105	0.028	0.078	0.419	0.023	0.149	0.043	0.168
Al2O3	15.1	16.0	15.4	20.9	14.1	15.0	14.3	15.0	15.2
Fe2O3	1.25	0.03	0.40	1.84	2.15	0.12	0.30	0.28	0.21
FeO	0.78	0.41	0.32	3.22	1.54	0.16	0.64	0.20	0.38
MnO	0.02	0.01	0.01	0.07	0.03	0.01	0.01	0.00	0.01
MgO	0.8	0.1	0.2	1.3	1.2	0.2	0.3	0.2	0.0
CaO	3.54	2.32	1.45	6.85	4.18	1.00	0.66	0.74	0.85
Na2O	4.9	3.9	4.5	5.2	3.7	2.5	2.3	2.8	1.1
K2O	0.48	2.96	2.76	0.96	0.86	7.41	6.79	6.50	4.08
P2O5	0.034	0.018	0.003	0.260	0.008	0.043	0.032	0.012	0.006
H2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CO2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
TOTAL	100.2	100.6	101.2	97.2	97.8	101.8	101.7	99.1	96.4

TRACE ELEMENTS IN PARTS PER MILLION

Cr	8	9	6	32	58	7	7	7	8
Ni	6	5	0	98	179	1	3	0	4
Cu	107	22	59	58	0	14	23	10	30
Zn	12	8	0	50	0	0	0	0	0
Rb	4	59	47	8	15	252	154	113	12
Sr	260	608	277	587	163	71	289	225	469
Yt	0	3	0	0	0	9	5	3	0
Zr	189	58	12	27	27	183	264	126	30
Nb	1	3	1	0	0	5	2	1	0
Ba	244	1185	564	294	399	1026	1991	833	414
La	67	18	0	34	7	0	55	10	3
Ce	84	21	0	61	0	0	102	29	0
Pb	14	28	13	19	7	39	24	33	9
Th	20	6	7	0	0	5	1	12	0

C.I.P.W. NORMS

Q	33.9	35.3	35.4	5.1	34.3	28.4	34.1	30.4	52.4
C	0.2	2.3	2.3	-	-	2.1	2.0	2.0	7.8
or	2.8	17.4	16.1	5.8	5.2	42.0	39.4	38.8	25.0
ab	41.4	32.7	37.7	45.1	32.4	20.9	19.1	24.1	9.4
an	17.3	11.3	7.1	31.9	19.8	4.6	3.0	3.6	4.3
nf	-	-	-	-	-	-	-	-	-
di	-	-	-	1.1	1.1	-	-	-	-
hy	2.1	0.7	0.8	7.4	3.1	0.7	1.5	0.6	0.4
ol	-	-	-	-	-	-	-	-	-
mt	1.8	0.0	0.6	2.8	3.2	0.2	0.4	0.4	0.3
il	0.4	0.2	0.1	0.2	0.8	0.0	0.3	0.1	0.3
hm	-	-	-	-	-	-	-	-	-
ap	0.1	0.0	0.0	0.6	0.0	0.1	0.1	0.0	0.0

6 BIOTITE PEGMATITES; 3 OTHER PEGMATITES

ANALYSIS NUMBER -	3011	2115	2941	2942	2943	2945	2146	2944	2946
SAMPLE NUMBER -	64NE1A	7061	5573	5806	72210	56113	GG	7341C	4932A
SiO2	72.6	73.6	70.1	72.2	69.2	72.9	71.0	66.7	65.9
TiO2	0.050	0.120	0.380	0.120	0.340	0.180	0.180	0.160	0.380
Al2O3	16.2	15.6	16.1	15.4	16.1	14.8	15.9	17.4	15.7
Fe2O3	0.63	0.16	1.36	1.11	1.02	0.83	0.67	0.75	2.94
FeO	0.47	1.26	0.51	0.78	0.98	0.57	1.00	0.37	0.78
MnO	0.01	0.01	0.02	0.01	0.01	0.02	0.02	0.01	0.05
MgO	0.2	0.1	0.1	0.2	0.8	0.7	1.2	0.6	1.1
CaO	2.32	0.39	0.92	0.56	0.60	0.62	3.18	2.41	2.41
Na2O	6.2	2.7	3.9	3.2	2.7	2.7	4.6	6.8	5.5
K2O	1.11	7.20	5.70	6.40	7.75	6.82	1.31	2.31	2.51
P2O5	0.007	0.032	0.097	0.020	0.031	0.037	0.031	0.057	0.095
H2O	0.99	0.40	n.d.	n.d.	n.d.	n.d.	n.d.	1.25	1.41
CO2	1.01	0.10	n.d.	n.d.	n.d.	n.d.	n.d.	0.50	0.59
TOTAL	101.7	101.6	99.1	99.9	99.5	100.1	100.1	99.3	99.5

TRACE ELEMENTS IN PARTS PER MILLION

Cr	8	4	0	5	9	5	33	9	17
Ni	1	0	0	1	7	8	46	1	17
Cu	16	0	0	15	9	0	3	15	5
Zn	0	0	8	17	29	0	28	5	48
Rb	78	119	103	147	156	170	53	50	84
Sr	202	322	367	439	497	343	394	340	488
Yt	0	4	15	4	3	5	0	1	9
Zr	214	196	2093	174	114	129	90	139	197
Nb	3	0	3	3	7	5	0	5	6
Ba	1052	2307	2500	1383	2658	2012	413	951	546
La	12	98	9	0	93	0	7	29	
Ce	11	166	99	0	151	0	2	55	
Pb	16	9	28	19	19	24	22	11	11
Th	2	16	151	6	0	23	2	6	9

C.I.P.W. NORMS

Q	35.0	29.2	24.2	27.8	21.5	28.9	31.4	16.0	23.0
C	2.9	2.9	2.2	2.3	2.2	1.0	1.2	0.7	1.2
or	6.5	42.0	34.0	37.9	46.0	40.3	7.7	13.9	15.1
ab	40.1	21.1	32.9	26.8	22.0	22.5	38.6	52.3	40.3
an	11.4	1.7	3.0	2.7	2.8	2.8	15.6	11.8	11.6
nf	-	-	-	-	-	-	-	-	-
di	-	-	-	-	-	-	-	-	-
hy	0.7	2.3	0.2	0.8	2.3	1.9	4.1	1.6	2.9
ol	-	-	-	-	-	-	-	-	-
mt	0.9	0.2	0.6	1.6	1.5	1.2	0.0	0.8	1.6
il	0.1	0.2	0.7	0.2	0.7	0.3	0.3	0.3	0.7
hm	-	-	0.0	-	-	-	-	0.2	1.9
ap	0.0	0.1	0.2	0.1	0.1	0.1	0.1	0.1	0.2

QUARTZOSE PEGMATITE; FLINTY CRUSH ROCK; EPIDOTIC CRUSH ROCK

ANALYSIS NUMBER -	3010	2939	2940
SAMPLE NUMBER -	4962B	4024	5577
SiO2	85.7	61.2	49.6
TiO2	0.020	0.600	0.280
Al2O3	8.3	14.9	15.4
Fe2O3	0.22	5.25	10.70
FeO	0.38	2.61	0.73
MnO	0.00	0.08	0.07
MgO	0.2	3.7	0.3
CaO	0.26	3.87	21.25
Na2O	4.4	2.9	0.0
K2O	1.15	1.75	0.46
P2O5	0.000	0.103	0.072
H2O	n.d.	2.00	1.93
CO2	n.d.	1.00	1.10
TOTAL	100.6	100.1	101.8

TRACE ELEMENTS IN PARTS PER MILLION

Cr	8	321	28
Ni	2	109	11
Rb	35	56	9
Sr	100	338	979
Yt	0	10	8
Zr	32	136	57
Nb	0	3	3
Ba	427	492	116
La	0	24	10
Ce	0	38	28
Pb	14	19	12
Th	0	0	4

C.I.P.W. NORMS

Q	54.9	33.3	23.4
C	-	3.9	-
or	6.8	10.6	2.7
ab	35.0	12.5	-
an	-	18.9	47.5
nf	-	-	-
di	1.1	-	1.8
hy	0.6	9.4	-
ol	-	-	-
mt	0.0	7.1	1.7
il	0.0	1.2	0.5
hm	-	0.5	9.5
ap	-	0.3	0.2

5 DIOPSIDE GNEISSES 4 EPIDOTE CHLORITE GNEISSES

ANALYSIS NUMBER -	2114	2150	2151	2154	2155	2928	2929	2930	2931
SAMPLE NUMBER -	7321I	7321D	7321E	7321H	7321J	5286	5551	5321F	5573
SiO2	59.4	65.1	55.0	62.0	62.4	64.7	71.1	71.5	67.5
TiO2	0.130	0.170	0.220	0.420	0.220	0.460	0.170	0.250	0.209
Al2O3	12.2	13.0	13.1	12.8	12.6	16.5	15.3	15.3	15.9
Fe2O3	1.13	2.07	1.37	1.43	1.56	1.99	0.78	1.13	1.17
FeO	3.80	0.94	2.12	1.75	1.57	1.40	0.71	0.89	0.96
MnO	0.16	0.14	0.22	0.13	0.17	0.04	0.01	0.02	0.04
MgO	6.6	5.4	7.2	5.9	7.1	1.2	0.4	0.3	1.1
CaO	7.63	5.62	14.07	6.17	7.20	3.30	0.70	0.88	1.39
Na2O	2.3	3.4	3.1	1.9	2.5	4.5	2.9	4.0	4.6
K2O	3.94	1.18	1.68	4.41	3.15	2.75	6.31	4.50	3.85
P2O5	0.070	0.051	0.096	0.108	0.066	0.126	0.073	0.066	0.135
H2O	2.51	1.92	1.42	1.42	1.52	0.49	0.73	0.50	2.16
CO2	0.06	0.24	0.71	0.71	0.31	0.21	0.12	0.00	0.06
TOTAL	99.0	99.2	99.8	99.1	100.4	97.8	99.3	99.4	99.1
TRACE ELEMENTS IN PARTS PER MILLION									
Cr	30	27	35	68	31	12	4	11	23
Ni	20	17	15	31	20	6	3	2	21
Rb	95	34	61	87	84	77	150	152	124
Sr	104	69	137	101	116	579	394	172	343
Yt	6	14	8	14	9	6	4	11	6
Zr	125	96	141	199	125	203	251	294	108
Nb	<5	5	<5	7	5	5	5	11	6
Ba	770	235	261	925	518	834	2690	675	1189
La	10	11	14	24	26	55	142	53	6
Ce	27	32	33	47	57	84	222	86	10
Pb	15	4	13	17	8	25	21	22	17
Th	8	6	8	9	9	5	19	9	0
C.I.P.W. NORMS									
Q	8.4	25.8	4.7	20.5	16.3	21.0	29.2	28.9	22.9
C	-	-	-	-	-	1.0	2.8	2.4	2.2
or	23.9	7.2	10.0	26.7	18.8	16.7	37.8	26.9	23.5
ab	19.3	26.4	18.3	8.0	17.6	36.2	23.8	34.4	39.3
an	11.9	18.9	21.3	18.1	15.9	15.0	3.0	3.0	6.2
nf	-	-	-	-	-	-	-	-	-
di	21.3	7.3	38.4	9.8	15.5	-	-	-	-
hy	13.0	10.3	2.9	11.9	12.1	3.3	1.4	1.1	3.3
ol	-	-	-	-	-	-	-	-	-
mt	1.7	3.1	2.0	2.1	2.3	2.0	1.2	1.7	1.8
il	0.3	0.3	0.4	0.8	0.4	0.9	0.3	0.5	0.4
hm	-	0.0	-	-	-	-	-	-	-
ap	0.2	0.1	0.2	0.3	0.2	0.3	0.2	0.2	0.3

EPIDOTE CHLORITE GNEISSES

ANALYSIS NUMBER -	2932	2933	2934	2935	2936	2937
SAMPLE NUMBER -	56110B	56110A	56116A	56116B	56116D	7481
SiO2	68.5	71.9	53.8	60.1	54.0	62.5
TiO2	0.246	0.185	0.876	0.773	0.858	0.752
Al2O3	15.5	15.6	14.9	15.1	14.0	14.2
Fe2O3	1.58	1.05	4.90	4.67	5.20	4.04
FeO	0.74	0.58	2.87	1.05	3.29	2.26
MnO	0.04	0.02	0.08	0.06	0.12	0.07
MgO	1.0	0.6	3.7	1.6	3.6	2.6
CaO	2.17	1.40	7.52	7.38	5.63	2.50
Na2O	4.1	3.4	2.9	4.4	3.4	4.3
K2O	3.65	4.79	1.49	1.07	1.98	1.47
P2O5	0.090	0.052	0.394	0.263	0.236	0.144
H2O	0.92	0.37	2.57	1.79	2.26	2.65
CO2	0.27	0.18	0.38	1.24	0.83	0.93
TOTAL	98.9	100.2	97.3	99.5	97.3	98.4

TRACE ELEMENTS IN PARTS PER MILLION

Cr	16	14	39	10	27	95
Ni	9	7	30	9	24	58
Rb	106	117	43	32	72	43
Sr	584	625	429	503	346	353
Yt	5	4	16	14	14	6
Zr	156	125	254	171	171	150
Nb	5	0	8	6	7	10
Ba	901	1294	414	292	473	512
La	24	13	74	36	44	33
Ce	42	27	135	68	89	62
Pb	20	17	11	9	10	11
Th	2	2	8	0	6	9

C.I.P.W. NORMS

Q	27.8	31.4	16.8	25.7	17.8	32.3
C	1.8	2.8	-	-	-	3.7
or	22.0	28.3	9.4	6.5	12.4	9.1
ab	31.9	26.8	21.1	22.7	20.1	26.1
an	10.4	6.6	27.5	26.8	26.5	11.0
nf	-	-	-	-	-	-
di	-	-	7.4	6.9	1.2	-
hy	2.6	1.6	6.5	0.9	9.5	6.7
ol	-	-	-	-	-	-
mt	1.8	1.4	7.4	1.4	8.0	5.6
il	0.5	0.4	1.8	1.5	1.7	1.5
hm	0.3	0.1	0.1	3.8	-	0.4
ap	0.2	0.1	0.0	0.6	0.6	0.4

V

ANTHOPHYLLITE GNEISSES

ANALYSIS NUMBER -	2102	2103	2104	2105
SAMPLE NUMBER -	70815B	70815C	70815D	70815E
SiO2	50.5	50.8	49.2	55.5
TiO2	1.490	1.230	1.020	0.670
Al2O3	12.1	13.6	13.9	12.9
Fe2O3	7.38	6.33	7.43	5.54
FeO	8.47	10.05	5.27	6.50
MnO	0.22	0.24	0.13	0.16
MgO	7.9	7.4	8.7	6.8
CaO	2.41	2.33	3.78	2.58
Na2O	4.3	4.2	4.1	5.9
K2O	0.16	0.12	0.45	0.17
P2O5	0.177	0.141	0.299	0.066
H2O	1.23	1.42	4.04	0.92
CO2	1.75	0.22	1.80	0.32
TOTAL	98.0	98.2	100.1	98.0

TRACE ELEMENTS IN PARTS PER MILLION

Cr	109	121	140	130
Ni	56	113	129	45
Rb	6	4	21	3
Sr	120	126	181	177
Yt	21	18	18	11
Zr	135	120	182	168
Nb	9	7	9	7
Ba	158	143	322	144
La	28	22	37	40
Ce	54	55	83	189
Pb	13	9	10	25
Th	4	4	10	13

C.I.P.W. NORMS

Q	20.2	6.8	17.8	5.4
C	5.2	3.3	4.9	-
or	0.0	0.7	2.8	1.0
ab	15.9	34.3	13.0	47.4
an	11.1	10.0	17.5	10.6
nf	-	-	-	-
di	-	-	-	1.8
hy	27.9	31.1	24.7	23.3
ol	-	-	-	-
mt	11.0	9.4	11.2	8.3
il	2.9	2.4	2.0	1.3
hm	-	-	-	-
ap	0.4	0.3	0.7	0.2

GARNET BIOTITE GNEISSES

ANALYSIS NUMBER -	2101	2106	2107	2108	2136	2156	2157	2158	2159
SAMPLE NUMBER -	70815A	PLK	PLA1	PLA2	52880	PLGR	7108	57810	T1D
SiO2	61.6	56.8	60.9	59.8	64.0	56.8	61.5	52.5	61.3
TiO2	0.680	1.120	0.730	0.770	0.420	1.120	0.680	1.760	1.490
Al2O3	16.0	14.0	14.9	14.0	14.7	14.1	15.0	12.4	8.5
Fe2O3	3.02	7.39	3.00	2.41	2.53	5.15	2.57	6.62	5.04
FeO	2.20	10.00	4.13	5.01	2.80	12.44	4.54	8.07	11.96
MnO	0.07	0.29	0.11	0.12	0.20	0.31	0.11	0.13	0.29
MgO	2.5	2.9	2.9	2.7	3.5	3.4	4.2	5.8	1.1
CaO	4.26	3.26	5.59	5.03	3.58	3.31	3.62	2.51	6.60
Na2O	4.0	2.4	2.5	2.7	3.9	2.6	3.2	2.9	2.3
K2O	1.45	1.21	1.68	2.74	1.81	1.16	2.30	3.19	1.00
P2O5	0.248	0.100	0.159	0.178	0.088	0.096	0.144	0.073	0.347
H2O	1.09	1.48	1.81	2.07	0.64	1.36	1.36	2.01	1.34
CO2	0.57	0.52	0.85	0.93	0.22	0.10	0.64	0.99	0.12
TOTAL	97.7	101.5	99.2	98.7	100.3	101.0	99.9	98.8	101.4

TRACE ELEMENTS IN PARTS PER MILLION

Cr	81	695	109	102	218	720	198	186	228
Ni	40	178	73	70	104	174	101	98	104
Rb	72	29	69	91	135	28	108	66	12
Sr	588	101	229	173	370	97	319	44	35
Yt	17	17	17	22	14	19	22	15	17
Zr	230	56	174	178	119	58	142	140	93
Nb	5	1	6	3	19	1	11	9	7
Ba	712	282	491	475	440	260	424	587	234
La	45	9	16	21	15	7	59	27	14
Ce	93	4	37	27	29	0	115	35	30
Pb	12	24	17	24	32	23	29	14	12
Th	13	0	1	8	10	0	54	6	0

C.I.P.W. NORMS

Q	25.8	26.3	30.0	24.7	23.3	18.1	23.8	15.2	27.6
C	2.1	4.2	1.2	1.0	0.5	3.0	2.6	2.1	-
or	8.9	7.1	10.2	16.6	10.8	6.8	13.8	19.5	5.9
ab	28.2	14.5	11.1	12.1	31.0	20.8	19.5	13.2	17.7
an	20.2	15.5	27.4	24.4	17.4	15.7	17.3	12.4	10.9
nf	-	-	-	-	-	-	-	-	-
di	-	-	-	-	-	-	-	-	17.1
hy	6.0	18.1	11.6	13.3	11.7	25.6	16.0	21.7	9.7
ol	-	-	-	-	-	-	-	-	-
mt	4.5	10.7	4.5	3.6	3.7	7.4	3.8	9.9	7.3
il	1.3	2.1	1.4	1.5	0.8	2.1	1.3	3.4	2.8
hm	-	-	-	-	-	-	-	-	-
ap	0.6	0.2	0.4	0.4	0.2	0.2	0.4	0.2	0.8

5 GARNET BIOTITE GNEISSES; 1 CALC SILICATE ROCK

ANALYSIS NUMBER -	2850	2867	2998	3004	3005	2137
SAMPLE NUMBER -	4965A	5301S	78SE1B	78SW6A	64SW5A	56910I
SiO2	63.4	50.7	59.5	60.0	60.4	47.6
TiO2	0.540	1.370	0.700	3.870	1.020	0.000
Al2O3	12.1	15.3	13.3	12.2	18.6	31.9
Fe2O3	2.97	7.00	5.53	5.05	2.17	2.69
FeO	5.04	14.00	6.34	9.23	5.73	1.42
MnO	0.15	0.38	0.08	0.11	0.05	0.06
MgO	4.3	3.7	4.0	1.7	3.5	1.2
CaO	4.97	3.25	3.58	3.64	2.10	7.34
Na2O	2.0	2.4	3.1	3.1	3.0	2.2
K2O	0.99	0.83	2.33	0.61	2.62	5.02
P2O5	0.144	0.067	0.171	0.102	0.021	0.015
H2O	1.65	1.81	0.73	0.82	0.48	n.d.
CO2	1.00	1.20	0.27	0.18	0.32	n.d.
TOTAL	99.3	101.0	99.7	101.6	100.0	99.0

TRACE ELEMENTS IN PARTS PER MILLION

Cr	271	1015	244	195	137	44
Ni	95	171	92	49	66	53
Rb	33	20	52	9	112	365
Sr	296	95	442	270	80	437
Yt	15	19	11	12	11	14
Zr	114	63	121	84	185	171
Nb	6	2	5	38	9	6
Ba	293	7	831	31	897	1485
La	15	231	17	71	29	0
Ce	36	0	24	117	56	16
Pb	31	30	18	18	8	8
Th	5	0	0	16	8	7

C.I.P.W. NORMS

Q	37.4	22.6	18.9	28.8	22.5	-
C	1.4	7.4	0.2	0.5	7.8	9.6
or	5.0	4.9	13.9	3.6	15.6	29.8
ab	5.3	6.3	23.5	23.9	21.9	16.3
an	24.3	15.7	16.8	17.3	10.3	36.5
nf	-	-	-	-	-	1.3
di	-	-	-	-	-	-
hy	17.4	27.4	16.2	10.8	16.0	-
ol	-	-	-	-	-	2.5
mt	4.4	10.1	8.1	7.3	3.2	3.9
il	1.0	2.6	1.3	7.3	1.0	-
hm	-	-	-	-	-	-
ap	0.4	0.2	0.4	0.2	0.1	0.0

MAFIC GRANOFELSES

ANALYSIS NUMBER -	2991	2992	2993	2994	2995	2996	2997	2999	3002
SAMPLE NUMBER -	64SW3A	64SW1A	64SW2A	64SW2B	78NW1B	78NW1C	64SW4	78SE1D	78NW1D
SiO2	49.1	51.0	49.7	50.9	49.3	48.7	48.9	50.9	47.1
TiO2	3.550	0.720	1.400	0.700	1.050	1.100	1.560	1.200	0.740
Al2O3	12.7	18.2	12.0	16.9	14.8	15.3	13.7	13.4	13.1
Fe2O3	5.42	4.27	6.61	4.77	4.52	3.22	5.57	2.87	3.63
FeO	8.27	4.03	7.42	3.51	7.24	9.78	8.97	7.58	10.47
MnO	0.22	0.12	0.21	0.10	0.17	0.25	0.20	0.29	0.23
MgO	4.5	4.1	5.7	4.4	6.3	4.4	6.7	4.9	6.4
CaO	8.05	7.28	11.34	7.69	8.53	10.57	8.89	11.10	12.72
Na2O	3.1	4.7	2.7	4.5	3.8	3.4	3.3	3.3	2.5
K2O	2.34	1.53	0.40	1.48	1.06	0.42	0.91	0.77	0.42
P2O5	0.414	0.388	0.141	0.463	0.083	0.068	0.140	0.096	0.066
H2O	1.23	1.39	1.80	2.40	2.53	1.90	0.72	2.05	1.56
CO2	0.41	0.71	0.10	0.60	0.50	0.40	0.28	0.56	0.52
TOTAL	99.2	99.4	99.6	98.4	99.9	99.5	99.9	99.0	99.5

TRACE ELEMENTS IN PARTS PER MILLION

Cr	103	29	143	16	88	141	155	148	120
Ni	74	26	67	23	117	49	103	85	72
Cu	21	0	83	65	5	55	0	4	6
Zn	63	98	78	89	90	95	95	121	96
Rb	57	15	7	17	10	7	5	5	11
Sr	175	1229	132	1384	408	134	91	232	98
Yt	31	15	20	9	12	28	15	21	14
Zr	189	64	89	33	73	51	71	75	38
Nb	12	4	3	3	6	0	5	3	0
Ba	2013	865	188	1074	418	168	167	235	159
La	21	36	9	28	14	0	6	11	2
Ce	47	80	21	66	25	5	16	33	14
Pb	15	14	18	16	14	13	9	15	2
Th	8	0	3	0	0	0	1	1	1

C.I.P.W. NORMS

Q	5.1	3.9	5.7	3.9	-	0.3	0.1	4.8	-
C	-	-	-	-	-	-	-	-	-
or	14.1	9.2	2.4	9.1	6.4	2.5	5.4	4.7	2.5
ab	21.3	31.8	22.2	32.1	27.3	24.8	24.9	21.9	15.1
an	16.0	29.2	20.6	26.3	23.8	28.2	21.7	23.7	27.3
nf	-	-	-	-	-	-	-	-	-
di	17.0	4.0	29.0	8.1	15.5	20.9	17.7	26.6	30.3
hy	8.6	11.6	6.9	9.2	16.7	15.1	18.0	9.0	15.2
ol	-	-	-	-	0.2	-	-	-	1.3
mt	8.0	6.3	9.8	7.2	6.7	4.8	8.1	4.3	5.4
il	6.9	1.4	2.7	1.4	2.1	2.1	2.0	2.3	1.4
hm	-	-	-	-	-	-	-	-	-
ap	1.0	0.9	0.3	1.1	0.2	0.2	0.3	0.2	0.2

1 MAFIC GRANOFELS; 8 GARNET BIOTITE GRANOFELSES

ANALYSIS NUMBER -	3003	2112	2119	2120	2121	2123	2124	2125	2848
SAMPLE NUMBER -	78SW6B	PLB	5724	5727A	5784	5301Z	5301Y	5301X	PLMA
SiO2	51.0	60.3	61.1	61.2	60.6	52.6	55.8	52.1	61.6
TiO2	0.980	0.663	0.659	0.570	0.571	1.137	0.800	0.718	0.550
Al2O3	13.4	14.3	14.9	14.4	14.8	17.6	14.6	16.7	15.3
Fe2O3	1.04	1.83	2.98	3.71	2.96	3.58	3.27	2.00	3.08
FeO	11.61	5.72	3.01	2.65	3.70	6.45	5.29	5.35	3.29
MnO	0.14	0.11	0.08	0.09	0.10	0.25	0.12	0.29	0.09
MgO	5.8	4.8	3.8	3.6	3.0	3.2	4.1	2.4	3.9
CaO	9.04	3.54	3.59	4.57	3.90	5.89	6.96	12.20	5.21
Na2O	3.1	2.0	3.7	3.4	3.8	3.5	3.4	4.2	3.0
K2O	0.56	2.31	2.41	2.15	2.15	2.36	1.39	1.16	1.45
P2O5	0.078	0.185	0.111	0.143	0.143	0.145	0.186	0.170	0.134
H2O	1.95	2.27	2.54	2.40	2.35	2.36	2.05	1.90	1.62
CO2	0.35	0.10	0.56	0.30	0.55	0.44	0.65	0.40	0.78
TOTAL	99.1	99.1	99.5	99.1	99.6	99.6	99.2	99.6	99.9

TRACE ELEMENTS IN PARTS PER MILLION

Cr	211	294	148	203	224	152	115	83	239
Ni	94	133	88	104	115	76	67	60	96
Rb	7	113	109	107	117	122	52	30	68
Sr	177	227	469	267	335	122	162	201	414
Yt	11	18	10	12	17	24	22	19	15
Zr	90	147	99	142	142	174	149	137	141
Nb	4	5	8	5	8	7	4	5	5
Ba	190	428	646	372	471	419	263	248	579
La	31	15	24	15	18	18	13	8	14
Ce	61	34	55	35	42	46	27	35	30
Pb	19	38	34	28	62	21	23	23	31
Th	6	0	12	3	4	3	0	0	2

C.I.P.W. NORMS

Q	1.0	17.6	21.7	21.6	20.1	6.6	15.4	1.2	27.5
C	-	1.2	1.3	-	0.9	-	-	-	1.4
or	3.4	14.1	14.7	13.1	13.1	14.4	8.5	7.0	8.7
ab	22.5	24.7	25.3	25.0	25.0	25.3	21.8	31.4	16.5
an	23.0	16.9	17.6	20.3	18.9	28.9	25.4	26.5	25.4
nf	-	-	-	-	-	-	-	-	-
di	18.1	-	-	1.7	-	0.1	7.4	27.5	-
hy	25.5	20.8	12.0	9.5	13.8	15.8	13.0	-	12.6
ol	-	-	-	-	-	-	-	-	-
mt	1.6	2.7	4.5	5.6	4.4	5.3	4.9	2.0	4.5
il	1.9	1.3	1.3	1.1	1.1	2.2	1.6	1.4	1.1
hm	-	-	-	-	-	-	-	-	-
ap	0.2	0.5	0.3	0.4	0.4	0.4	0.5	0.4	0.3

GARNET BIOTITE GRANOFELSES

ANALYSIS NUMBER -	2849	2855	2856	2857	2858	2869	3000	3001
SAMPLE NUMBER -	T5A	4007E	4007F	7488	70212	7491	78SE2B	74NE3B
SiO2	61.2	57.1	61.3	57.1	57.4	59.7	54.8	56.3
TiO2	0.680	0.690	0.610	1.000	0.642	0.590	0.860	1.590
Al2O3	14.0	15.5	13.7	15.0	14.7	14.9	9.7	17.4
Fe2O3	3.73	4.87	3.18	3.82	3.82	3.12	6.10	2.29
FeO	3.19	3.00	3.92	5.38	3.83	3.30	16.87	8.11
MnO	0.10	0.10	0.11	0.22	0.10	0.09	0.28	0.19
MgO	4.5	4.9	4.2	3.5	5.2	4.1	3.6	2.0
CaO	3.80	3.90	4.99	4.02	2.86	4.88	3.62	2.53
Na2O	3.0	3.4	2.0	2.8	3.1	3.5	1.3	4.3
K2O	2.59	2.48	1.85	2.86	2.62	1.70	0.87	3.63
P2O5	0.158	0.142	0.127	0.194	0.098	0.123	0.279	0.042
H2O	1.52	2.48	2.53	3.15	3.51	2.69	1.02	1.26
CO2	0.56	1.00	0.12	0.62	1.48	0.44	0.22	0.23
TOTAL	100.1	99.5	99.6	99.7	99.6	99.1	99.4	99.8

TRACE ELEMENTS IN PARTS PER MILLION

Cr	231	292	286	114	257	193	253	120
Ni	104	107	64	51	124	93	77	38
Rb	112	73	59	111	116	68	20	68
Sr	343	388	370	154	317	495	72	81
Yt	15	13	13	26	12	18	65	16
Zr	156	130	133	173	143	155	64	167
Nb	5	6	4	5	8	4	7	6
Ba	523	611	477	457	502	543	374	571
La	19	11	0	17	17	17	28	36
Ce	39	19	3	33	41	41	49	59
Pb	68	35	24	27	21	12	15	7
Th	3	10	0	4	5	2	2	3

C.I.P.W. NORMS

Q	23.5	21.4	21.1	19.7	27.3	19.7	22.6	6.5
C	1.0	2.0	-	1.9	5.4	-	1.3	2.5
or	15.5	15.1	11.3	17.5	16.2	10.4	5.2	21.8
ab	19.2	17.1	24.6	17.0	9.3	25.3	8.3	33.7
an	18.1	18.0	19.9	19.3	14.1	23.5	16.4	12.5
nf	-	-	-	-	-	-	-	-
di	-	-	3.8	-	-	0.6	-	-
hy	13.2	13.0	12.8	14.8	16.7	13.2	34.4	15.0
ol	-	-	-	-	-	-	-	-
mt	5.5	7.3	4.8	5.7	5.7	4.7	8.0	3.4
il	1.3	1.3	1.2	1.0	1.3	1.2	1.7	3.1
hm	-	-	-	-	-	-	-	-
ap	0.4	0.4	0.3	0.5	0.2	0.3	0.7	0.1

HORNBLENDE BIOTITE GRANOFELSES

ANALYSIS NUMBER -	2118	2122	2126	2127	2128	2129	2134	2853	2854
SAMPLE NUMBER -	7472	7042	5301W	5301V	PLE	PLD	4964A	708180	B3
SiO2	59.4	63.6	58.8	52.0	67.2	68.8	57.8	61.9	54.8
TiO2	0.660	0.480	0.630	0.690	0.550	0.620	0.690	0.510	0.690
Al2O3	14.5	14.6	15.7	13.8	13.7	13.4	15.3	15.0	15.5
Fe2O3	3.80	2.92	3.27	1.14	2.29	2.23	3.52	3.58	4.54
FeO	3.23	2.84	2.83	5.29	3.11	2.98	4.14	2.00	4.07
MnO	0.11	0.07	0.07	0.34	0.07	0.09	0.11	0.08	0.13
MgO	3.7	3.5	3.8	3.6	2.9	2.3	4.3	2.8	4.6
CaO	5.79	3.69	5.23	17.82	2.89	3.55	4.20	3.95	4.19
Na2O	3.4	3.4	3.6	1.5	3.9	4.2	3.8	3.0	3.3
K2O	1.61	2.04	1.93	1.08	2.08	1.34	2.49	1.95	2.68
P2O5	0.154	0.117	0.115	0.195	0.154	0.128	0.177	0.139	0.152
H2O	2.76	2.44	2.18	1.61	1.56	0.89	2.33	2.11	2.75
CO2	0.20	0.46	0.45	0.19	0.44	0.11	0.37	1.39	1.04
TOTAL	99.3	100.2	98.7	99.3	100.9	100.5	99.3	100.3	98.6

TRACE ELEMENTS IN PARTS PER MILLION

Cr	72	196	63	33	125	159	199	140	342
Ni	55	86	55	39	61	74	97	73	169
Rb	67	83	79	32	166	116	137	106	136
Sr	349	480	258	81	156	180	349	513	388
Yt	21	13	23	17	12	22	18	18	12
Zr	185	149	183	125	172	199	143	151	170
Nb	5	5	5	2	9	7	5	8	10
Ba	397	590	341	435	312	226	565	510	491
La	12	15	14	8	15	44	16	22	34
Ce	44	28	30	32	42	70	43	41	68
Pb	19	21	34	16	23	31	24	30	27
Th	8	3	0	2	10	8	8	1	8

C.I.P.W. NORMS

Q	18.8	26.7	17.9	6.7	28.6	28.0	13.4	30.0	18.1
C	-	1.5	-	-	1.1	-	0.1	3.7	2.5
or	9.8	12.3	11.8	6.5	12.4	7.9	15.2	11.7	16.5
ab	27.6	23.9	25.9	10.6	28.2	34.1	28.2	17.5	15.0
an	21.4	17.0	24.8	29.7	13.4	14.6	20.3	19.0	20.7
nf	-	-	-	-	-	-	-	-	-
di	5.8	-	1.0	35.8	-	1.8	-	-	-
hy	8.7	11.0	11.0	-	10.5	7.6	14.9	7.0	15.0
ol	-	-	-	-	-	-	-	-	-
mt	5.7	4.3	4.9	1.7	3.3	3.2	5.3	5.3	6.9
il	1.3	0.9	1.2	1.3	1.0	1.2	1.3	0.0	1.4
hm	-	-	-	-	-	-	-	-	-
ap	0.4	0.3	0.3	0.5	0.4	0.3	0.4	0.3	0.4

3 HORNBLENDE BIOTITE GRANOFELSES; 3 DIOPSIDE GRANOFELSES

ANALYSIS NUMBER -	2868	2938	3027	2147	2148	2149
SAMPLE NUMBER -	56914A	7042B	T5C	7321A	7321B	7321C
SiO2	61.9	58.9	83.1	61.8	62.4	59.4
TiO2	0.620	0.653	0.150	0.530	0.430	0.600
Al2O3	16.3	13.6	8.3	14.9	16.1	15.5
Fe2O3	2.79	5.24	1.17	1.71	1.79	2.43
FeO	2.19	3.24	0.46	1.79	1.59	1.72
MnO	0.06	0.10	0.07	0.08	0.04	0.05
MgO	2.4	3.5	4.3	5.2	3.7	5.8
CaO	3.37	3.96	2.41	3.44	1.78	1.96
Na2O	4.4	2.3	0.8	2.3	2.1	2.5
K2O	2.63	3.81	1.49	5.18	7.91	3.79
P2O5	0.161	0.118	0.391	0.178	0.113	0.192
H2O	2.54	2.10	n.d.	1.74	1.42	1.82
CO2	0.53	1.00	n.d.	0.17	0.17	0.54
TOTAL	99.0	98.5	101.8	99.8	99.6	96.2

TRACE ELEMENTS IN PARTS PER MILLION

Cr	102	198	23	71	58	70
Ni	55	85	27	38	31	31
Rb	138	81	121	138	157	118
Sr	541	323	31	110	92	133
Yt	11	11	9	21	0	10
Zr	374	124	130	319	168	412
Nb	11	3	4	10	5	10
Ba	862	885	601	882	1525	959
La	44	8	11	56	20	79
Ce	80	24	20	102	44	135
Pb	36	17	12	33	14	3
Th	2	5	3	15	12	18

C.I.P.W. NORMS

Q	20.2	26.5	60.5	15.4	11.0	23.8
C	1.7	1.2	1.8	0.1	1.5	5.8
or	15.0	23.3	8.6	31.5	47.6	23.7
ab	32.1	7.6	6.9	17.8	16.3	15.4
an	16.1	19.6	9.2	16.4	8.2	8.0
nf	-	-	-	-	-	-
di	-	-	-	-	-	-
hy	6.0	9.9	10.4	14.5	10.3	15.5
ol	-	-	-	-	-	-
mt	4.1	7.9	1.2	2.6	2.6	3.7
il	1.2	1.3	0.3	1.0	0.8	1.2
hm	-	-	0.3	-	-	-
ap	0.4	0.3	0.9	0.4	0.3	0.5

MAP 1

Geology of Coll, Argyll

N

1km.

Traigh Bousd

Bagh an Trailleich

Cliad Bay

Rudha Ard

Hogh Bay

Feall Bay

Crossapoll Bay

Loch Breachacha

Loch Eatharna

Caolas Ban

Calgary Point

Traigh Halum

Eilean Bhorcnuil

Ben Hogh

Ben Feall

Fasagh

Cnoc Shorbidh

Meall nan Uan

Bagh na Caille

Sonsdale

Loch Fada

Feosdlum

Loch Anlaimh

Eilean Orrisay

Port na Luing

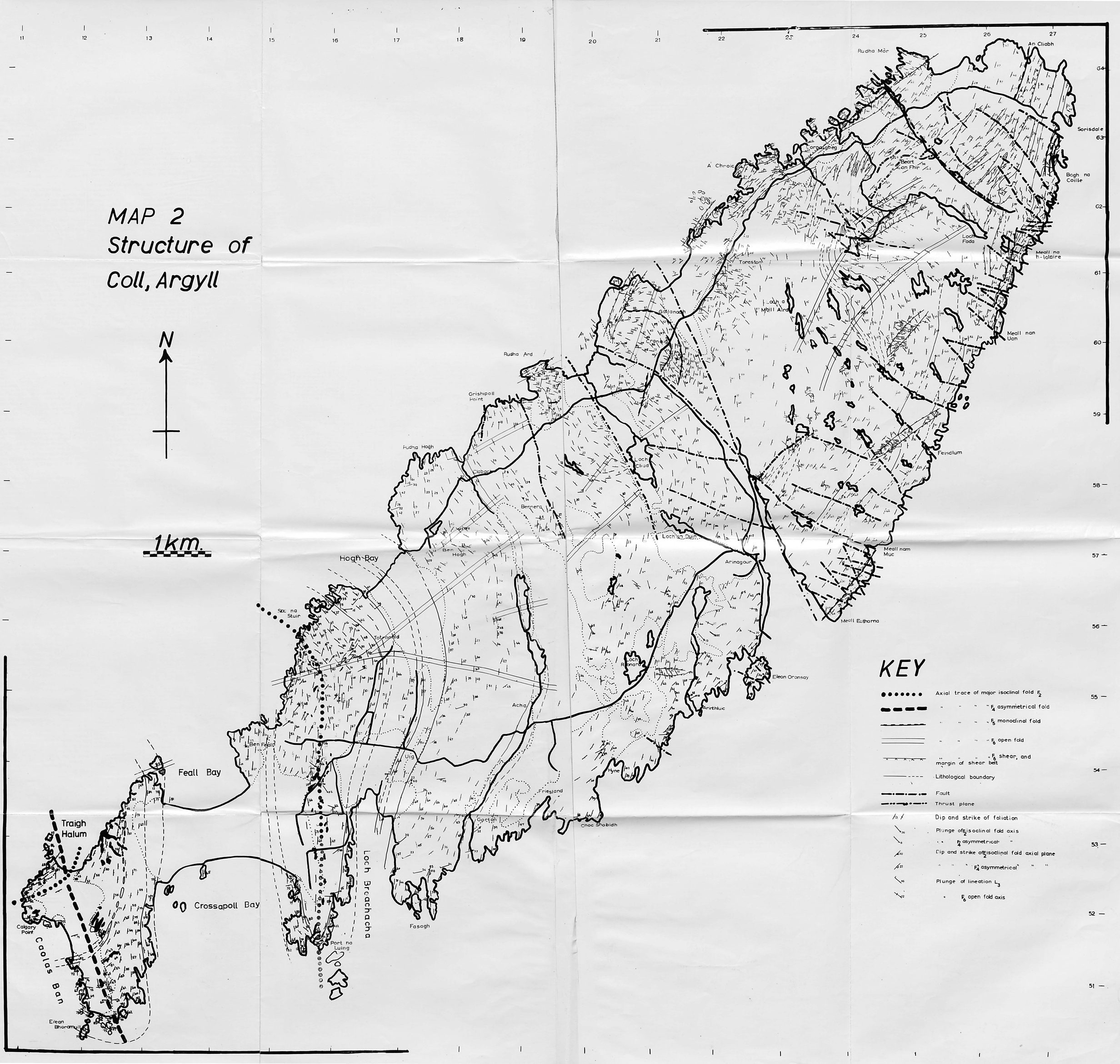
KEY

- Metabasite
- Agmatite
- Ultramafic rock
- Granofels
- Concordant granite
- Quartzose pegmatite
- Biotite-pegmatite
- Migmatitic gneiss
- Nebulitic gneiss
- Sheared gneiss
- Quartzite
- Calc-silicate rock
- Marble
- Garnet-biotite-gneiss
- Anthophyllite-gneiss
- Diopside-gneiss
- Superficial deposit

MAP 2
Structure of Coll, Argyll

N

1km.



KEY

- Axial trace of major isoclinal fold F_2
- " " " " F_4 asymmetrical fold
- " " " " F_5 monoclinal fold
- " " " " F_6 open fold
- " " " " F_6 shear, and margin of shear belt
- Lithological boundary
- Fault
- Thrust plane
- Dip and strike of foliation
- Plunge of F_2 isoclinal fold axis
- " " F_4 asymmetrical " "
- Dip and strike of F_2 isoclinal fold axial plane
- " " " F_4 asymmetrical " " "
- Plunge of lineation L_3
- " F_6 open fold axis

MAP 3

50

45

40

Fhadamuill

Vaul Bay

Salum

Balephetr

Bhasapoll

Rudha Chraiginis

Beirn Hough

Gott Bay

Brock

Rudha Nead

N

Scarinish

Eilean Ghreasamail

Traigh nan Gilean

Hynish Bay

Balephuil

Ben Hynish

Cean a' Mhara

Hynish

Geological Map: Lewisian of Tiree

KEY

As Map 1

Hypersthene-gneiss

Strike of vertical foliation

m. Magnetite band

1mile
1km.

95 0 05 10